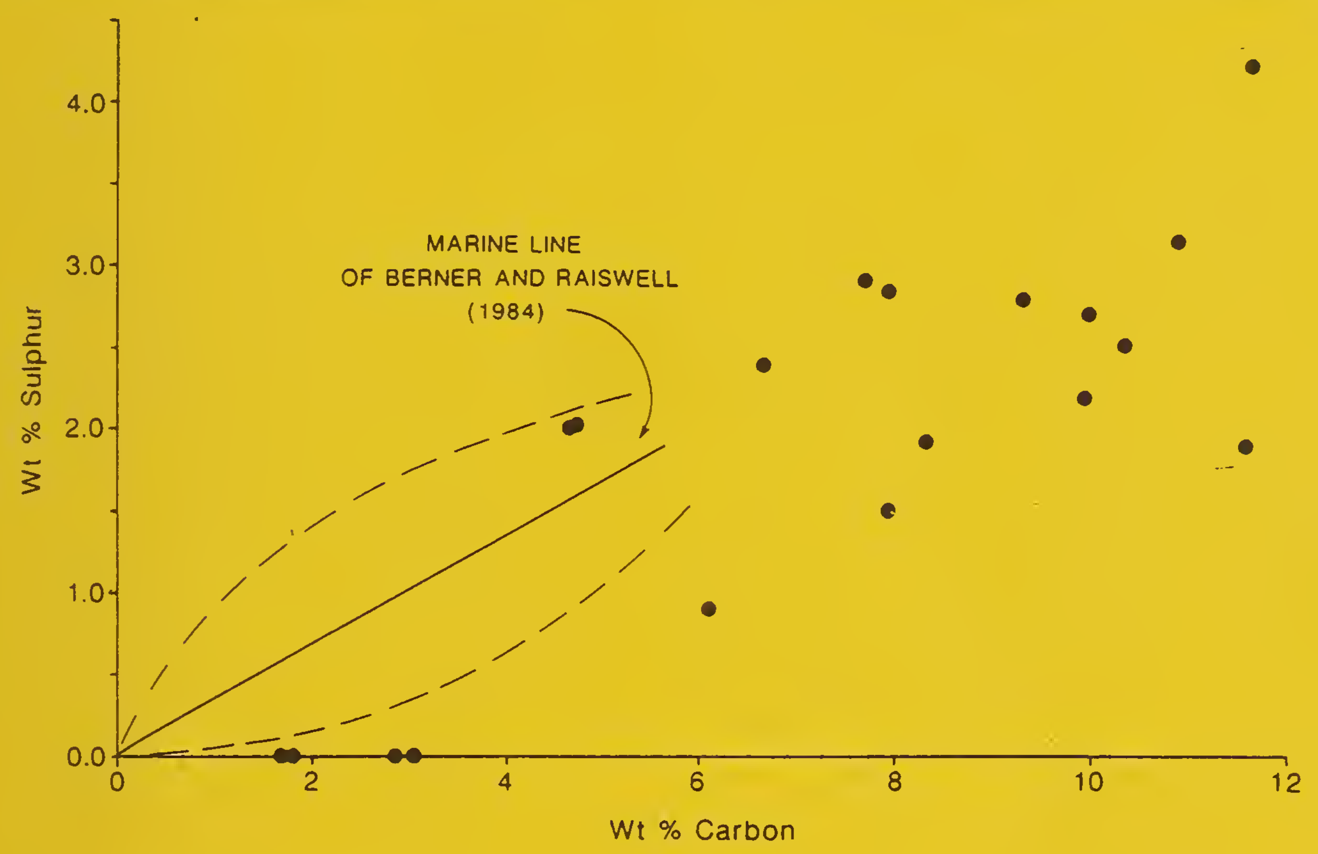
BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

(FOR ARTICLES ON THE GEOLOGY OF EAST ANGLIA)

NO.45 for 1995



PUBLISHED 1996

CONTENTS INCLUDE

Pre-Glacial and Glacial Sands and Gravels, How Hill, Norfolk

Carbon and Sulphur Geochemistry, West Runton Freshwater Bed

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No. 45 (for 1995) Published 1996

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EDITORIAL

Bulletin No. 45 concentrates on two aspects of Pleistocene geology in Norfolk. The paper by Rose *et al*. describes pre-glacial sands and sands and gravels at How Hill. These sediments were deposited in coastal environments but were influenced by sediment inputs from major rivers which were flowing into the sea close by. The paper by Hannam *et al.* is a geochemical and mineralogical study of the West Runton Freshwater Bed. The results show that the upper part of the bed has been altered (oxidised) by groundwater some time after deposition. In addition, the clay mineral smectite was found to be abundant in the bed and may indicate input of volcanic dust.

I am short of material for future issues of the Bulletin and welcome the submission of papers on any aspect of East Anglian geology. I hope to publish Bulletins 46 and 47 in 1997 to bring the publication schedule up to date. This will, however, depend on the submission of suitable articles

INSTRUCTIONS TO AUTHORS

If possible, contributors should submit manuscripts as word-processor print out accompanied by a disk copy. We can handle most word-processing formats although PC Word, WordPerfect or ASCII files are preferred. In addition we accept typewritten copy and will consider legible handwritten material.

It is important that the style of the paper, in terms of overall format, capitalisation, punctuation, etc. conforms as strictly as possible to that used in Vol. 41 of the Bulletin. Titles and first order headings should be capitalised, centred and in bold print. Second order headings should be centred, bold and lower case. Text should be 1½ line spaced. All measurements should be given in metric units.

References should be arranged alphabetically in the following style.

BALSON, P.S. & CAMERON, T.T.J. 1985. Quaternary mapping offshore East Anglia. *Modern Geology*, **9**, 221-239.

STEERS, J.A. 1960. Physiography and evolution: the physiography and evolution of Scolt Head Island. In: Steers, J.D. (ed.) *Scolt Head Island* (2nd ed.), 12-66, Heffer, Cambridge.

BLACK, R.M. 1988. *The Elements of Palaeontology*. 2nd Ed., Cambridge University Press, Cambridge. 404pp.

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The editors welcome original research papers, notes or comments, and review articles relevant to the geology of **East Anglia** as a whole, and do not restrict consideration to articles covering Norfolk alone. All papers are independently refereed by at least one reviewer.

PRE-GLACIAL AND GLACIAL QUATERNARY SEDIMENTS, HOW HILL NEAR LUDHAM, NORFOLK, ENGLAND

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ABSTRACT

Trial pit excavations at How Hill near Ludham, Norfolk, have revealed a succession of pre-glacial Quaternary sediments overlain by Anglian till. The pre-glacial deposits consist of sands, and sands and gravels. These are interpreted on the basis of their sedimentary structures, lithologies, particle size distribution and palaeocurrent directions, as having formed in coastal environments with a fluvial input. Palaeocurrents show NNW-SSE trending tidal current flow, associated with a contemporary coastline. Provenance-indicator lithologies show dominant sources of sediments from both the south and the north-west, by the Thames and a 'Northern rivers' system. One of the trial pits revealed a water-escape structure within the sorted sediments partially infilled with till, suggesting rapid dewatering of the sands and gravels after glaciation, possibly associated with the decay of permafrost. The absence of any lithologies diagnostic of the glaciogenic sediments of eastern England indicates a pre-glacial origin, and the association of sediments from a Kesgrave Group and 'Northern river's' source suggests that eastern Norfolk was a depositional focus for major rivers of pre-glacial Britain. Glaciation at the site is represented by a weathered till, known locally as Norwich Brickearth, which was deposited during the Anglian Stage.

Bull. geol. Soc. Norfolk (for 1995) 45, 3-28 (Published 1996)

INTRODUCTION

This paper presents the results of a study of three trial pits excavated through till and sands and gravels at How Hill near Ludham, eastern Norfolk. The results provide new evidence about the sand, and sand and gravel lithostratigraphic units of the region. The sedimentology of the sands indicates formation by tidal currents trending NNW-SSE, probably associated with a contemporary coastline, and the gravel units represent periodic channel sedimentation. The gravels have fractions derived from south-eastern and Midland England, the Pennines and also Wales, but without a glacially derived component, and are attributed to transport by major pre-glacial river systems of southern and midland Britain. An impressive wedge at one of the sites has deformed the upper part of the sorted sediments and has also disturbed the base of the overlying diamicton. This diamicton is weathered till, known locally as the Norwich Brickearth (Perrin et al., 1979). The wedge is interpreted as a water escape structure formed by rapid dewatering of the sorted sediments after the deposition of the glacial sediments and is attributed to high pore-water pressures generated by the decay of permafrost following glacierisation. The sands and gravels formed in the Early Pleistocene and the diamicton was deposited during the Anglian Stage of the Middle Pleistocene.

This work is the outcome of a collaborative research programme between the Department of Geography at Royal Holloway, University of London (RHUL) and the British Geological Survey (BGS) to study the Pleistocene gravels and diamictons in East Anglia. The work is targetted upon supporting the regional mapping programme which is part of the BGS long-term strategy (British Geological Survey, 1990) and an investigation of major fluvial and coastal sedimentary units that were deposited pre-glacially in eastern Britain (Rose, 1994; Hamblin and Moorlock, 1995).

The site recorded here is at Crome's Farm, How Hill near Ludham, Norfolk (TG 377 199) (Fig. 1). The area consists of low, undulating relief, with a height range of about 7m, rising above Crome's Broad which has a water surface level at about 0 m O.D. The site is located in the catchment of the river Ant. The regional geology is recorded on BGS maps as Chalk overlain by Palaeogene sediments and Crag which is capped by till (Cromer Till/ Norwich Brickearth) (1:63,360, Old Series Sheet 66NE, surveyed in 1875-1880, and 1:250,000 Sheet, East Anglia).

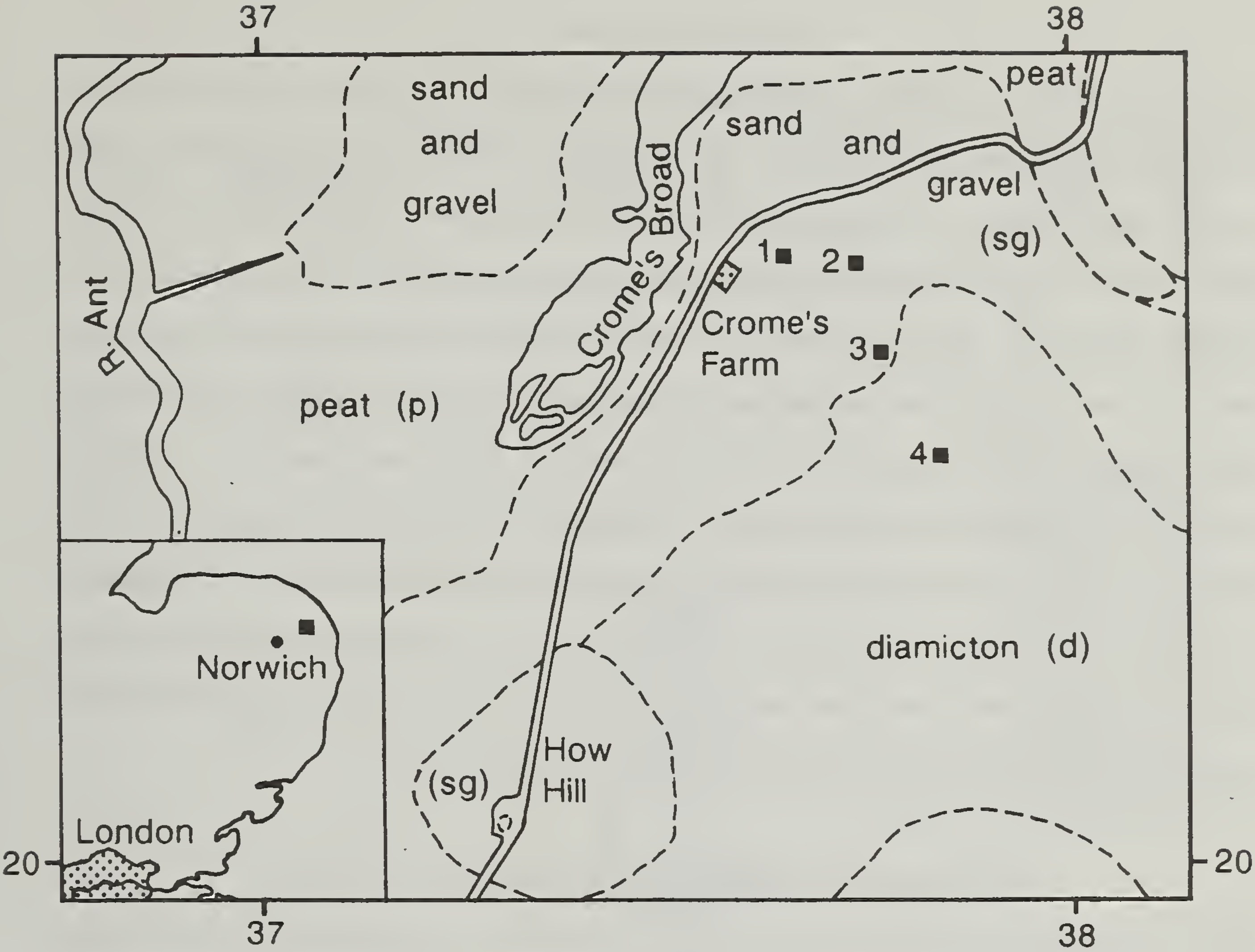

Fig. 1. Location of trial pits near How Hill, Ludham, Norfolk. The site position is shown in the inset. Scale and orientation are shown by 1 km national grid coordinates. The site is located in 100 km grid square TG. The geology of the site is shown by the surface lithologies mapped by the British Geological Survey.

Previous work on the area has been restricted because of the limited number of exposures. The site is 15 km ENE of Catton Brick Pit where Hey and Brenchley (1977) and Hey (1980) recognised Kesgrave Sands and Gravels, and 23-30 km SE of the north Norfolk coast between Weybourne and Trimingham where Green and McGregor (1990) recognised a complex sequence of pre-glacial gravels involving transport from the south and west. Samples from the diamicton in the area were analysed as part of the study on the Pre-Devensian tills of eastern England by Perrin et al. (1979), which showed that the glacial deposits are typical of weathered Cromer Till, known locally as Norwich Brickearth.

METHODOLOGY

Four trials pits were dug with a JCB excavator to a depth of about 4 m. The surface level of each pit was surveyed accurately to a local Ordnance Survey spot height shown on the 1:10,000 scale map. Pit 1 flooded, and does not contribute to this study. Facies logs were derived for each section including sedimentary structures and sediment colour (Munsell notation). Dip and azimuth of depositional cross-set structures in the sands were measured as palaeocurrent indicators, although it was not possible to obtain such measurements from the predominantly structureless sands and gravels. Seventeen samples were collected for particle size analysis (Table 1), clast lithological analysis (Table 2), and, in one case, palynological investigation (Quaternary and pre-Quaternary). A sample of diamicton was taken in a Kubiena box for micromorphological analysis, using the preparation procedures outined in Lee and Kemp (1992). All analyses were carried out in the Geography Department laboratories at RHUL.

SITE DESCRIPTIONS AND INTERPRETATIONS

The three trial pits are located along a north-south transect which rises to the south. Pit 2 has a surface elevation of 4.7 m and Pit 4, 8.3 m O.D. (Fig. 2). All positions given relate to dimensions shown on the figures.

Pit 2 (Fig. 3)

This pit was 3.90 m deep, and exposed interbedded sands and gravels overlain by c. 1 m of sandy diamictic colluvium. The sands and gravel units are structureless. Size distributions show modes in the 8-16 mm and the 250 μm-1 mm fractions (Fig 4, Table 1) characteristic of deposition, respectively, by traction and saltation (Visher, 1969). The intervening sand beds are well sorted with modes of up to 70% in the 250-500 μm fraction. These show well developed cross-set structures with palaeocurrent directions towards the north and south-west with, locally, herringbone patterns with an azimuthal variation of 95° (Fig. 5). Occasionally the depositional structures are dissected by channels. A 2 cm thick clay drape covers the base of one of these channels and occasional thin (<1cm) silty clay lenses cover sand beds.

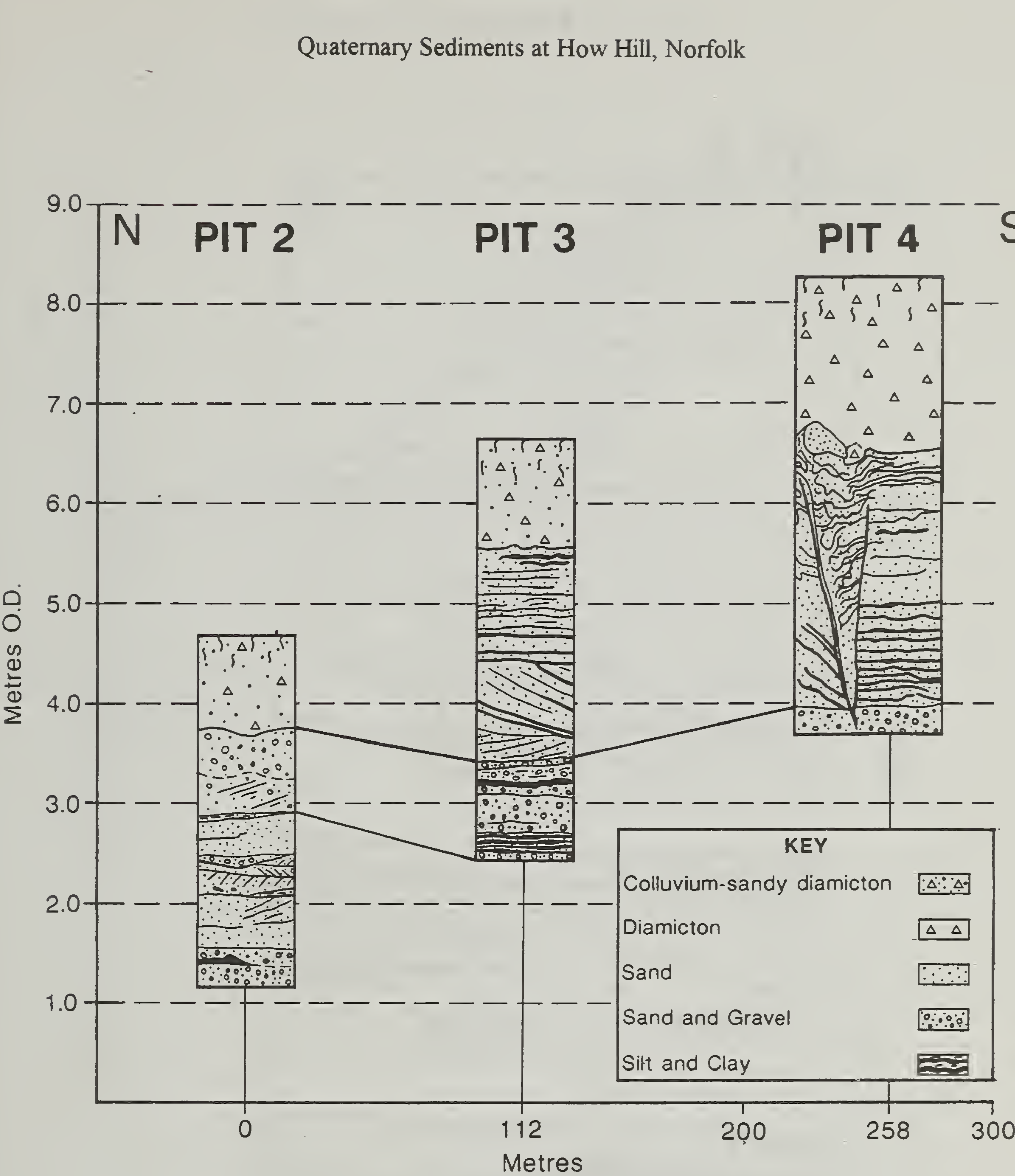

Fig. 2. Relationship of trial pit sections 2, 3 and 4 to one another and to topography. The location of these sites is shown on Fig. 1. A tentative correlation of one of the sand and gravel units is indicated.

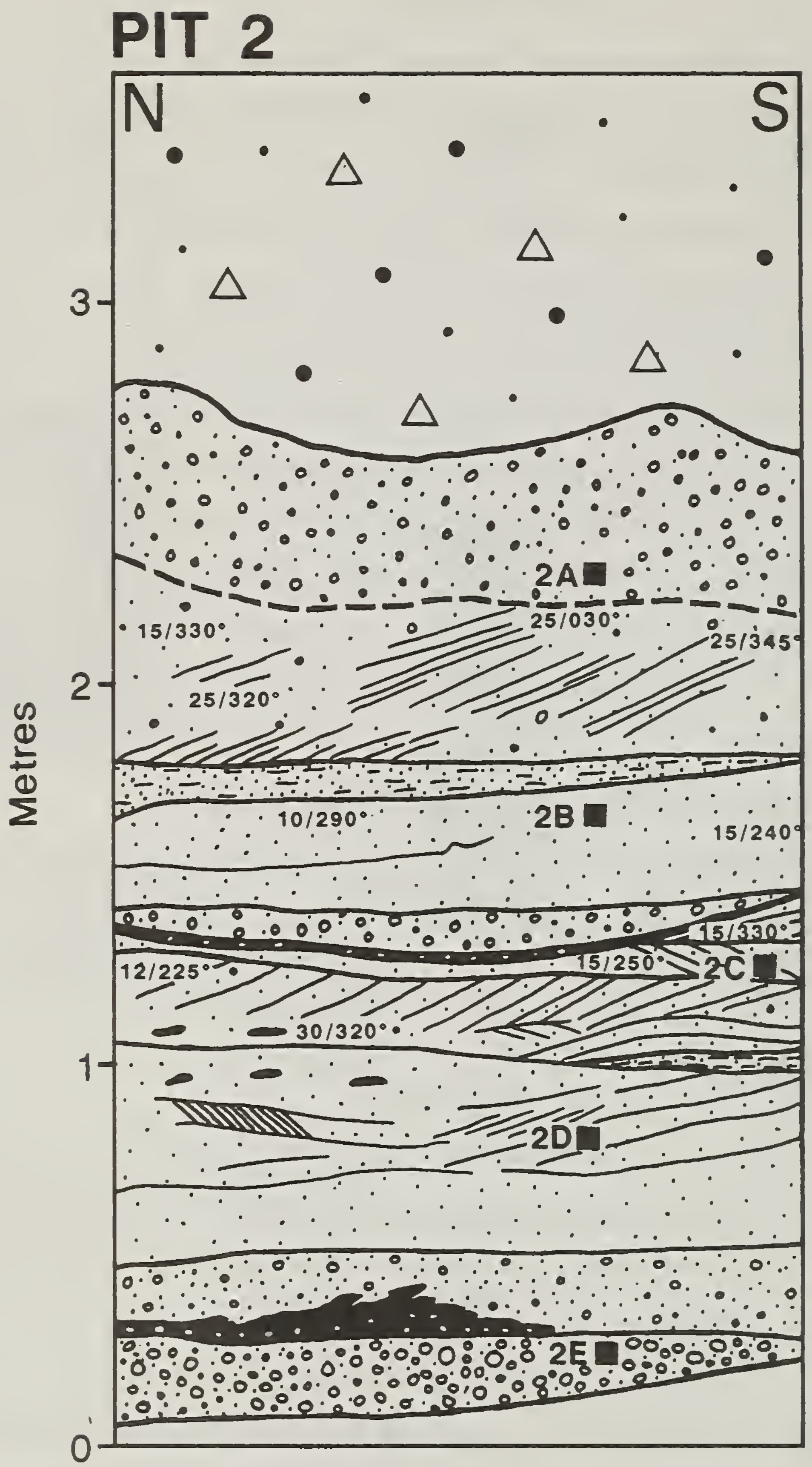

Fig. 3. Representation of sedimentary structures at Trial Pit 2, How Hill. This is a true scale drawing with dimensions shown at the margins. Locations of sample points for particle size and clast lithological analysis are indicated by a black square, along with sample number. Palaeocurrent dip and orientation are shown at the point of measurement. The key to symbols is shown in Fig. 2.

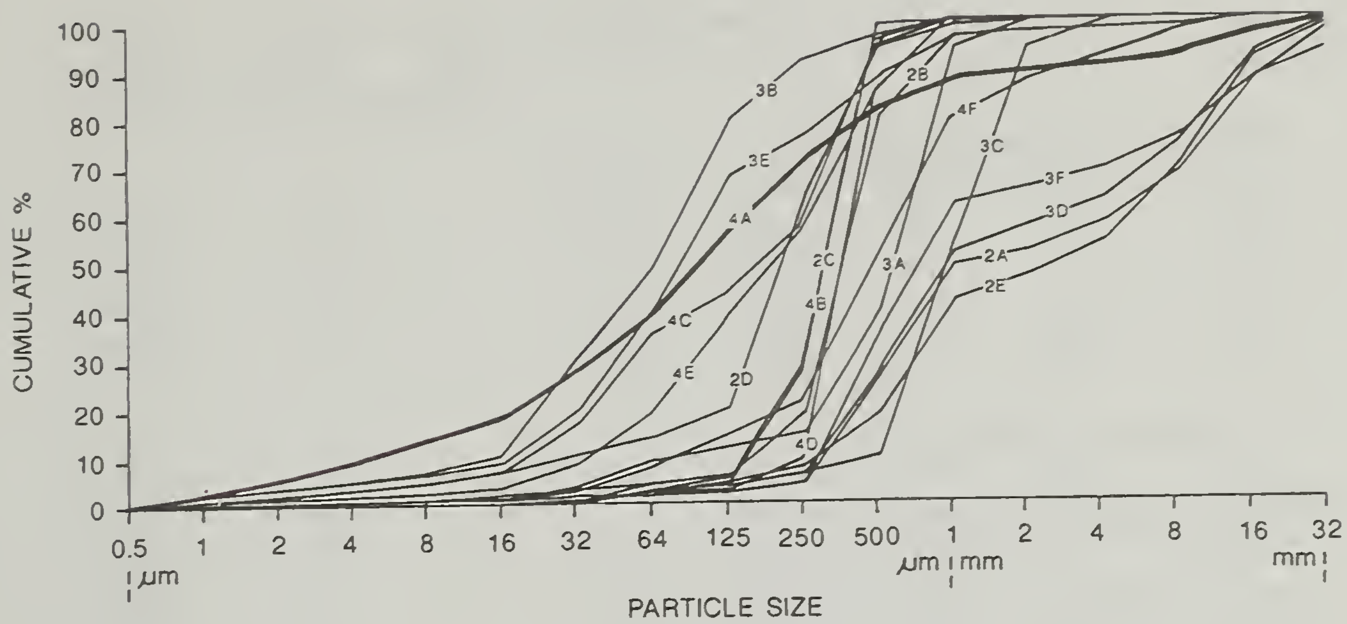

Fig. 4. Particle size distributions of all samples from sections exposed in trial pits at How Hill, near Ludham, Norfolk.

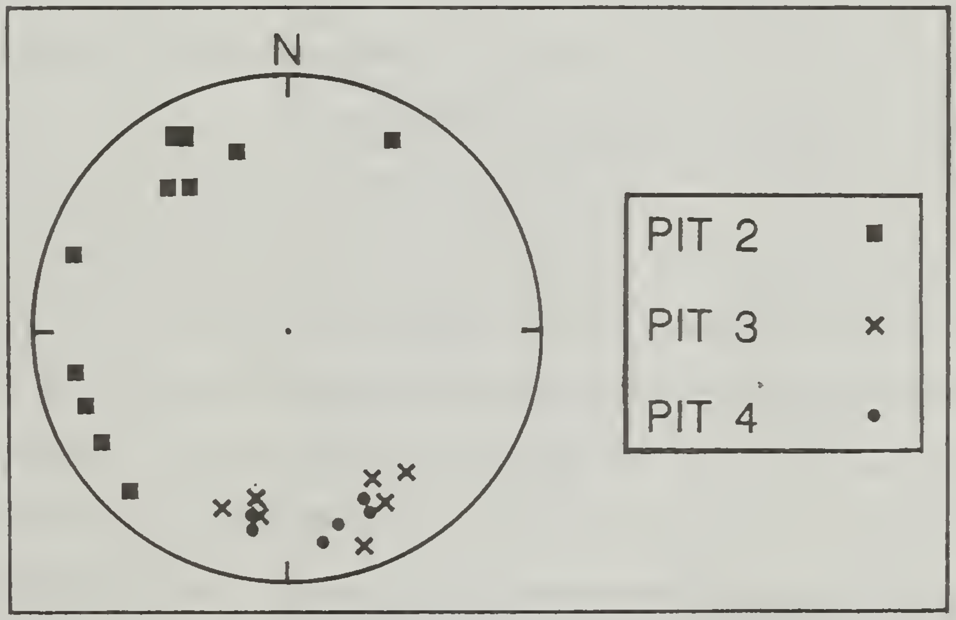

Fig. 5. Dip and orientation of palaeocurrent measurements from sections exposed in Trial Pits 2, 3 and 4 at How Hill, near Ludham, Norfolk. These measurements are taken on the slope of depositional cross-sets shown in Figs. 3, 4 and 5 or at the sides of Pit 4.

Table 1. Particle size composition of the sediments from How Hill, near Ludham, Norfolk.

Site No.	% Clay	% Silt	% Sand	%Gravel	Gravel mode (phi size)	Sand mode (phi size)	Total sample weight (kg)
Pit 2. Sands and sands and gravels							
2A	0.10	1.59	50.53	47.78	-3	+1	7.503
2B	0.41	3.82	93.49	2.28		+2	0.101
2C	-	2.70	97.30	-		+2	0.104
2D	2.49	11.44	85.76	0.31		+3	0.105
2E	0.31	1.96	44.85	52.88	-3	+1	11.803
Pit 3. Sands and sands and gravels							
3A	0.46	8.46	90.75	0.33		+1	0.101
3B	3.44	45.57	50.99	-		+4	0.090
3C	-	4.32	90.21	5.47		+1	0.103
3D	0.13	1.93	55.38	42 57	-3	+1	6.261
3E	3.35	36.71	57.58	2.36		+4	0.103
3F	0.08	1.46	64.02	34.44	-3	+2	11.160
Pit 4. Diamicton							
4A	5.94	33.69	49.78	10.59	-3	+4	6.707
Pit 4. Sands and sands and gravels							
4B	-	2.64	97.36	-		+2	0.106
4C	1.87	33.04	65.09	-		+2	0.040
4D	-	2.19	97.75	0.06		+2	0.103
4E	1.15	18.00	80.85	-		+2	0.102
4F	0.28	7.18	79.20	13.34	-1	+2	4.250

Pit 3 (Fig. 6)

This pit was 4.3 m deep and exposed a gravel unit at the base (0.0 - 1.1 m) with an interbedded silty fine sand, and bedded sands with interstratified silty clays above. It is capped by a sandy diamictic colluvium similar to that in Pit 2. The sorted sediments show colour banding related to the re-deposition of iron oxides. Most of the sandy gravels are structureless with size modes in the 8-16 mm and 250 µm - 1 mm fractions (Fig. 4, Table 1), again typical of transport and sedimentation by traction and saltation. At c. 0.9 m the gravel includes an irregular silty clay layer, probably reflecting load deformation (Visher, 1969). The silty fine sand shows a well developed mode in the 63-125 µm fraction and a long tail of coarser sands typical of fine-grained sedimentation where sand sized material has rained out of suspension due to rapid changes in the current regime. This explanation probably accounts for the occurrence of the thinner silty fine sand units elsewhere in the gravels.

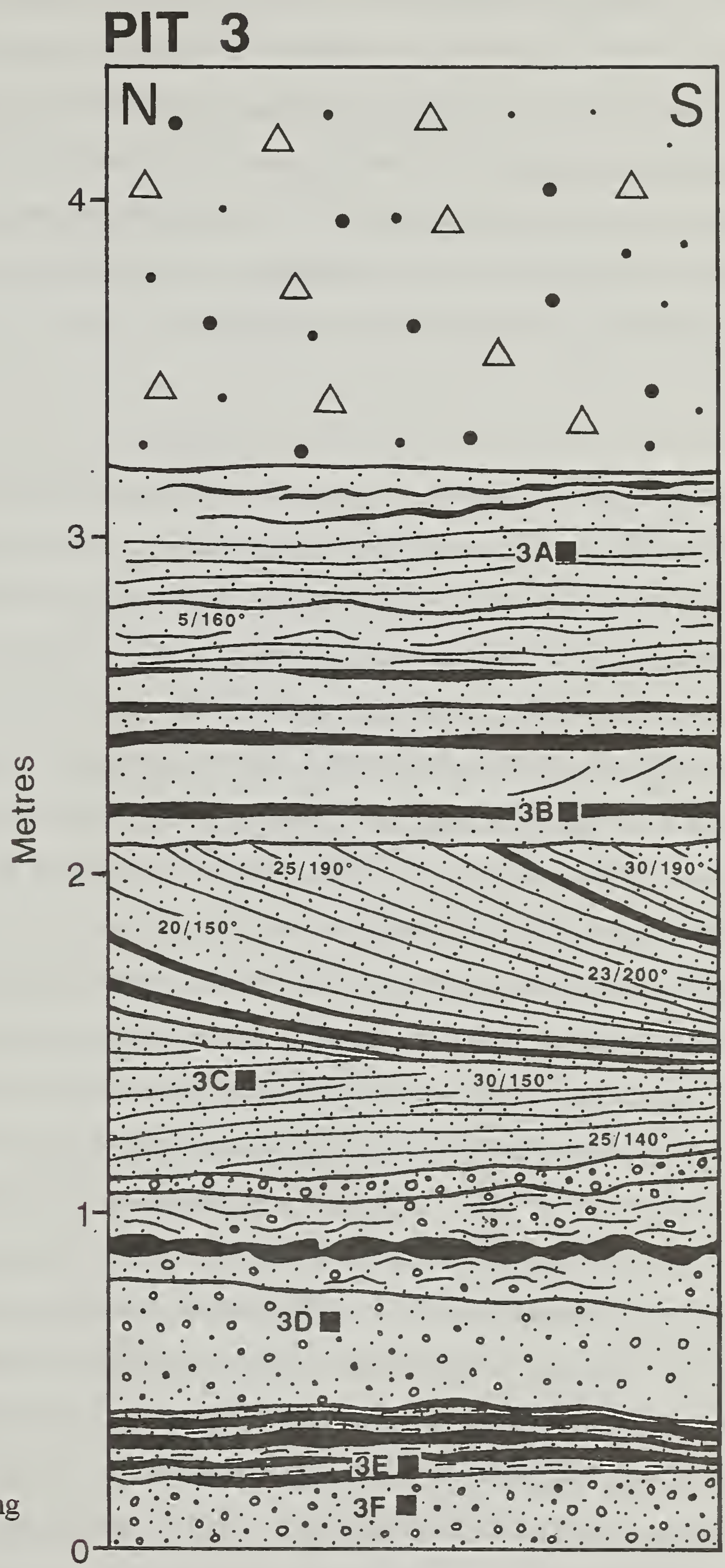

Fig. 6.
Representation of sedimentary structures at Trial Pit 3, How Hill. This is a true scale drawing with dimensions shown at the margins. Locations of sample points for particle size and clast lithological analysis are indicated by a black square, along with sample number. Palaeocurrent dip and orientation are shown at the point of measurement. The key to symbols is shown in Fig. 2.

The sands are interbedded with micaceous silty clays. They are moderately well sorted with a mode of up to 85% in the 500 μm - 2 mm fraction (Fig. 4, Table 1). Between 1.1 and 2.1 m they form large-scale cross-beds (amplitude up to 0.5 m) with a palaeocurrent direction towards the south (Figs 5 and 6). Between 2.1 and 3.2 m the sands are horizontally bedded with silty clay laminations and clay drapes over ripple laminations (flasers). These sands show a mode (55%) in the 500 μm -1 mm range typical of high flow regime flat-bed sedimentation.

Pit 4 (Fig. 7)

This pit was 4.5 m deep and exposed a gravel unit at the base, overlain by 2.4 m of horizontally bedded sands with silty sand laminae, and capped by 1.65 m of decalcified sandy diamicton. The sorted sediments are deformed by a wedge-shaped structure which extends into the lower part of the diamicton. The sands are dominated by modes in the 250-500 μm fraction (68% sample 4B, 75% sample 4D) (Fig. 4, Table 1) typical of a saltation population transported by uniform current flow. The fine laminae, between the sand beds and within depressions on the surfaces of the sand beds retain a 250-500 μm mode (29% sample 4E, 36% sample 4C) and a tail of silt size material (18% sample 4E, 33% sample 4C) which suggests that these units were deposited from saltation with a minor fraction raining out from suspension. Measurements on small-scale cross beds within the sands show flow directions towards the south-east (Fig. 5). The combination of sedimentary structure, size range, and flow direction indicators suggest that these sands were deposited during episodes of persistent current flow interrupted by lower energy events during which the fine suspended load settled out.

The diamicton is reddish brown (7.5YR 6/5), with a multimodal size distribution. Although dominated by the 63-125 μm fraction, minor modes exist in the 4-8 μm and 8-16 mm size ranges. With respect to these properties this deposit is typical of the Norwich Brickearth of Norfolk (Perrin et al., 1979) and can be compared with sample No 1876 (Ludham, TG 398 191), which contributed to the analysis used in Perrin et al. (1979), bearing in mind the different position within the weathering profile (2.5 m below surface for Sample No 1876, 1.45 m below surface in Pit 4).

Micromorphology of the diamicton shows a granular sediment with rounded, subrounded and angular sand and silt size grains in a clay matrix. The sand grains are

composed predominantly of quartz and flint fragments, with occasional micas, feldspars, glauconite, shale fragments and till pebbles (van der Meer, 1993). There is no evidence of weathering of any of the quartz and most of the other lithologies have sharp edges, but there is evidence of fretting at the margins of occasional feldspars and micas, and most glauconite and shale fragments have disintegrated. There is no evidence for 'ghosts' of weathered grains (such as limestones), but the matrix shows stress orientation with moderately birefringent clays, which may indicate that voids have subsequently been obscured. In all these respects the deposit fulfils the requirements of a weathered and re-organised till (van der Meer, 1993). Evidence for soil formation within the diamicton is spectacular with multi-layered void and grain coatings of clay, coarse clay and silty clay, and silty-clay cappings (Kemp, 1987; Whiteman and Kemp, 1990) and occasional calcite overgrowths (Kemp, 1985). Void and grain coatings are both undisturbed and disturbed and, together with the re-deposited calcite, show a complex history of soil formation. Together, the micromorphology confirms the interpretation that the deposit is a weathered till and confirm the designation as Norwich Brickearth (Perrin et al. 1979).

The wedge shaped structure is 2.67 m deep with a maximum width of 0.8 m at the top. Its boundaries are well-defined and continuous (Fig. 7a). The structures at the northern side show a zone with parallel boundaries c. 0.05 m wide (Fig. 7a). Except at the top, the material included in the wedge is of the same composition as that outside the structure. The structures within the wedge show a downward displacement at the south side, a horizontal configuration in the centre and upward adjacent to the parallel zone. At the top of the north side of the wedge, the sands penetrate upwards into the diamicton and at the south side a lobe of diamicton descends into the sand.

First appearances suggested an ice-wedge cast (Black, 1976), but with further analysis (P. Worsley, pers. comm., 1995), the autochthonous composition, the upturned structures on the north side, and collapse structures on the south side suggest that the feature formed by soft sediment deformation. This occurred due to water escaping upwards through the parallel zone on the north side which is, in effect, a pipe with material collapsing into the wedge-shaped void on the south side (Fig. 7b). The upward injection of sand into the diamicton, and collapse of the diamicton into the wedge means that this process took place after the diamicton was deposited indicating that the wedge formed after the site had been glaciated. The absence of glacial deformation at the

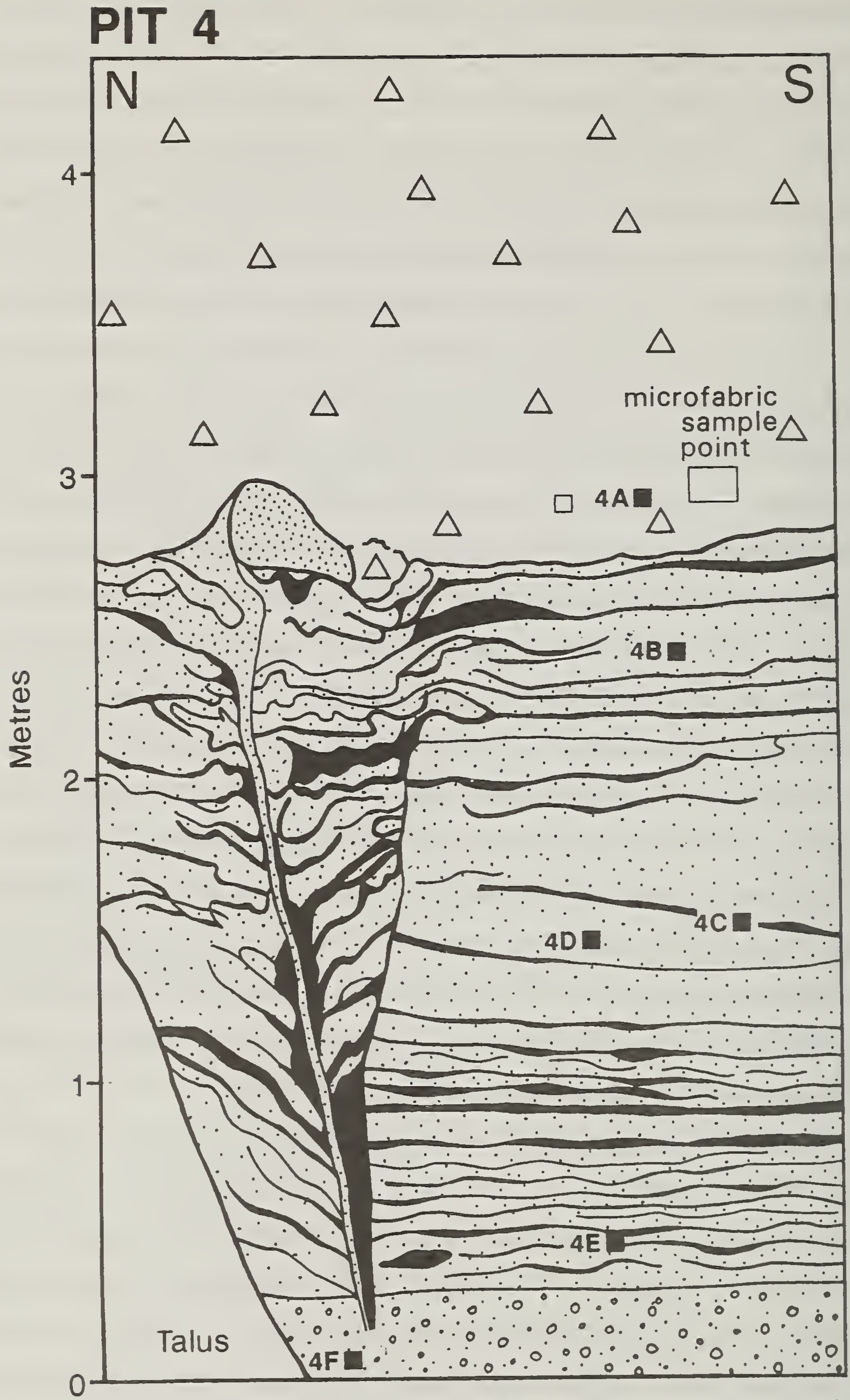

Fig. 7a. Representation of sedimentary structures at Trial Pit 4, How Hill. This is a true scale drawing with dimensions shown at the margins. Locations of sample points for particle size, and clast lithological analysis are indicated by a black square, along with sample number. The location of the sample taken for microfabric is shown by an open rectangle. The key to symbols is shown in Fig. 2.

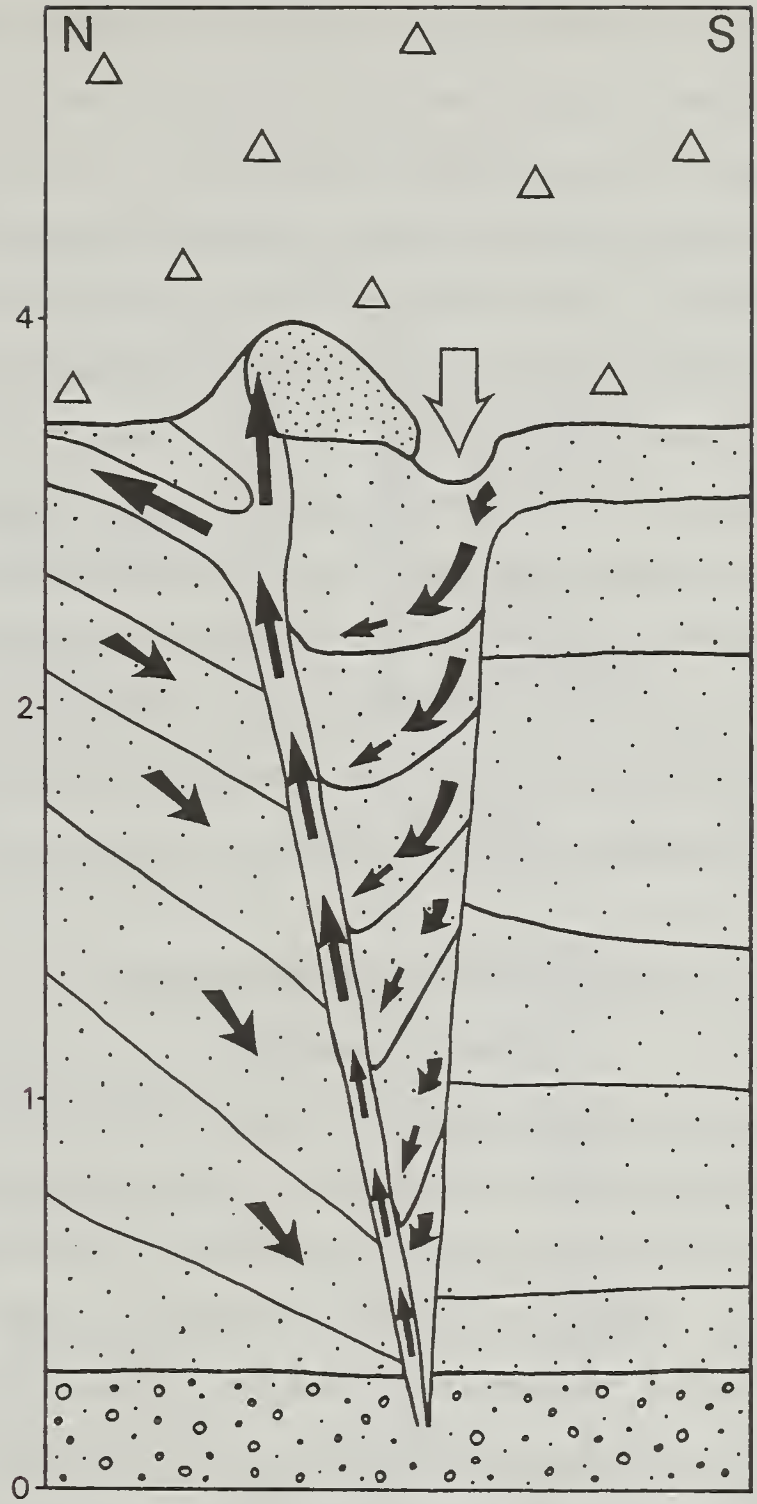

Fig. 7b. Cartoon to illustrate the mechanism proposed for the formation of the water-escape structure exposed in Pit 4. See text for detailed explanation.

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contact between the sand and the diamicton suggests that this process took place after ice movement had ceased. Thus the structure was formed by pore-waters escaping from the sand and sands and gravels. It is suggested that with the constraints outlined above, the most likely conditions for this process to take place are following the decay of permafrost, either due to geothermal heat trapped beneath glacier ice, or due to climatic amelioration.

PROCESSES OF FORMATION OF THE SORTED SEDIMENTS

The form, structure, palaeocurrent properties and size distribution of the sand units are typical of high energy dune and flat-bed sedimentation with intervening low-flow conditions. The silty sands reflect lower energy conditions with the silt and fine sand raining out of suspension. These processes, together with the occurrence of flaser bedding and bi-directional flow patterns, suggest tidal flow in a coastal environment.

The sands and gravel units are formed by traction and saltation typical of high energy current transport by rivers or tides. The relatively high percentages of gravel and the predominantly structureless form of the beds is typical of river sedimentation, although aggradation at the base of tidal current channels is equally possible, and the association with silt drapes and tidal sands favours this depositional environment.

The relationship between the three pits is shown in Fig. 2, which also shows elevations above OD and suggests a possible correlation of one of the gravel units.

LITHOLOGICAL COMPOSITION

Method of Analysis

The lithologies of four gravel and one diamicton samples were analysed at the 8-16 mm and 16-32 mm size ranges (Table 2). As the clasts are more abundant at the smaller size range and the range of lithologies is greater, interpretation of the sands and gravels is based on these results. Because of the relative scarcity of clasts, the diamicton is described in terms of the 8-32 mm fraction. Lithological groups were classified, as far as possible, according to their probable geographical provenance (Fig. 8). Thus, quartz, quartzites and schorl from the Triassic Kidderminster Formation of the West Midlands are grouped together. Similarly, chatter-marked flints from either the Westleton Beds of East Anglia or the Tertiary pebble beds of the London Basin, angular flints from the Cretaceous Chalk of southern and eastern England, and Greensand chert from Kent are also grouped together. Acid volcanic rocks from Wales, Carboniferous chert from the southern Pennines, Spilsby Sandstone from the southeast Lincolnshire region, and *Rhaxella* chert from north-east Yorkshire are identified separately. Clearly these relationships can be broken down by reworking during a complex transport history and this is considered in the following discussion.

Table 2a. Lithological content of the gravels and diamicton at How Hill, expressed as percentage of total sample number (n).

Site No	Size Range in mm	n	Carb. chert	Triassic qzt	Triassic qz	Triassic schorl	Triassic total	Jurassic Rhaxella chert	Cretaceous sst.	Cretaceous glauc. sst	Cretaceous Green-sand chert	Cretaceous flint	Cretaceous total	Tertiary/ Pleistocene chattermarked flint	Igneous	Unknown
Gravels																
2A	8-16	708	5.6	16.0	22.6	0.8	39.4	0.0	2.0	0.0	0.6	42.9	45.5	8.8	0.3	0.4
2E	8-16	1482	4.9	14.1	23.0	1.3	38.4	0.1	0.9	0.2	0.3	47.4	48.8	7.0	0.3	0.1
3D	8-16	671	6.0	7.1	17.5	1.5	26.1	0.6	2.1	0.0	2.1	57.3	61.5	5.1	0.0	0.6
3F	8-16	692	7.7	14.6	16.9	1.4	32.9	0.6	0.0	1.2	0.4	50.4	52.0	6.4	0.3	0.1
All	16-32	252	6.0	11.9	9.5	0.4	21.8	0.4	1.2	0.0	0.4	61.2	62.8	7.9	0.0	1.2
Diamicton																
4A	16-31	151	0.0	8.6	15.3	0.0	23.9	0.7	0.0	0.0	0.0	64.2	64.2	8.6	1.3	1.3

Table 2b. Distinctive lithological sub-fractions as percentage of sample number (n), and ratio values.

Site No	Size range in mm	n	Triassic white/ colourless quartzite	white/ colourless quartz	coloured quartzite	coloured quartz	Qtz:Qz ratio	Coloured: colourless qtz+qz ratio	Flint: qtz+qz +schorl ratio
Sands and Gravels									
2A	8-16	708	10.5	18.4	5.5	4.2	0.71	0.34	1.15
2E	8-16	1482	10.7	19.6	3.4	3.2	0.61	0.20	1.27
3D	8-16	671	6.1	15.6	1.0	1.9	0.41	0.13	2.36
3F	8-16	692	9.4	12.9	5.2	4.0	0.86	0.41	1.58
All	16-32	252	7.1	9.1	4.8	0.4	1.25	0.32	2.93
Diamicton									
4A	16-31	151	6.6	11.3	1.4	4.4	0.56	0.33	2.69

Description

The bulk of the gravel samples are composed of angular flint (57-43%) and quartz and quartzite (39-26%). The next most abundant lithologies are chatter-marked flints (9-6%) and Carboniferous chert (8-5%). Lithologies that indicate a specific provenance include *Rhaxella* chert (3 samples, 0.6-0.1%), glauconitic Spilsby Sandstone (2 samples, 1.2-0.2%), Greensand chert (all samples, 2.1-0.3%) and acid volcanic rocks from Wales (3 samples, 0.3%).

The diamicton sample differs from the sands and gravels in that it contains a higher proportion of angular flint (64.2%), a lower proportion of quartz and quartzite (24%), a roughly similar amount of chatter-marked flint, and a higher proportion of *Rhaxella* chert. The igneous rocks are not lithologically distinctive and all other indicator rocks (Carboniferous chert, glauconitic Spilsby Sandstone and Greensand chert) are absent.

Sample 3E was analysed for pollen but proved to be barren.

Provenance

The absence of any calcareous rocks in these samples, in a region that is close to chalk outcrops suggest that both the sands and gravels and the diamicton are decalcified throughout, although it is noted that the chalk in the area is 'almost ubiquitously soft' and comminution during current transport may be responsible for its absence from the sands and sands and gravels (B.M. Funnell, pers. comm. 1996).

The composition of the sands and gravels indicates either a remarkably wide provenance or major reworking of existing sediments that have a very wide provenance (Fig. 8). In either case material contributing to the gravels shows sources to the north and northwest (*Rhaxella* chert, glauconitic Spilsby Sandstone and Carboniferous chert), in the Midlands (Triassic quartz and quartzite), in southern England (Greensand chert) and in Wales (acid volcanic rocks). The chatter-marked flints cannot be differentiated as to whether they are derived from the Westleton Beds, the Bure Valley Beds of the Crag Group, or from the Tertiary (Palaeogene) pebble beds.

Assuming a fluvial origin for these sediments there are three possible drainage systems by which the gravels could have reached How Hill (Fig. 9). The Welsh acid volcanic rocks, Kentish Greensand chert, Triassic quartz and quartzite, angular flint and

chatter-marked flints are typical of the Kesgrave Sands and Gravels (Hey, 1976, 1980; Whiteman and Rose, 1992), and reflect transport by the River Thames at the time the 'great Thames' drained much of Wales and southern and Midland England (Rose, 1994). At this time the river Thames continued its course northwards through East Anglia to the Thames/ Rhine/ Eastern Rivers (North German Rivers) delta in the North Sea (Gibbard, 1988, Green and McGregor, 1990). The quartz and quartzite, glauconitic Spilsby Sandstone, and Carboniferous chert are typical of the Bytham Sands and Gravels (Rose, 1987, 1994) reflecting drainage by a major river from the west Midlands and the southern Pennines, but not including Wales. This river joined the region in central East Anglia (Rose, 1989, Figure 33). The third provenance is indicated by the *Rhaxella* chert from the Howardian Hills of Yorkshire (Hey, 1976), the Spilsby sandstone and the Carboniferous chert. In this case provenance is likely to be from a northern source, possibly involving rivers that drained the southern Pennines. This river system does not have a name, and will be known here, for convenience, as the 'Northern rivers' (Fig. 9). Based on geomorphological evidence of consequent river patterns (De Boer, 1974) the 'Northern rivers' are unlikely to have transported the *Rhaxella* chert directly to north-east Norfolk and coastal transport of materials, either carried down these northern rivers, or transported to the coast by glaciation, as suggested by Hey (1976), seems a more likely explanation.

Clearly there is a problem in that the Carboniferous chert and the Spilsby Sandstones could have been transported either by the Bytham river or the 'Northern rivers'. It is possible to attempt to solve this problem by reference to the character of the quartz and quartzite in these deposits (Table 2, Fig. 8). The quartz and quartzites in the Bytham Sands and Gravels are typically reddish brown and brown with only low percentages of white/ colourless lithology. In contrast the quartz and quartzites in the Kesgrave Sands and Gravels are dominantly white/ colourless irrespective of their age (Hey 1976, 1980) (Fig. 8b), as are those in the Carboniferous gritstones and Brassington Formation in the southern Pennines (Barke et al., 1920; Walsh et al., 1972), which are other possible sources of quartz and quartzites . A plot of these properties (Fig. 8b) shows that the sediment at How Hill is dominated by white/ colourless quartz and quartzite and is unlikely to be derived from the Bytham river, but is typical of the Thames sediments. Therefore, composition of the How Hill gravels suggests that they were

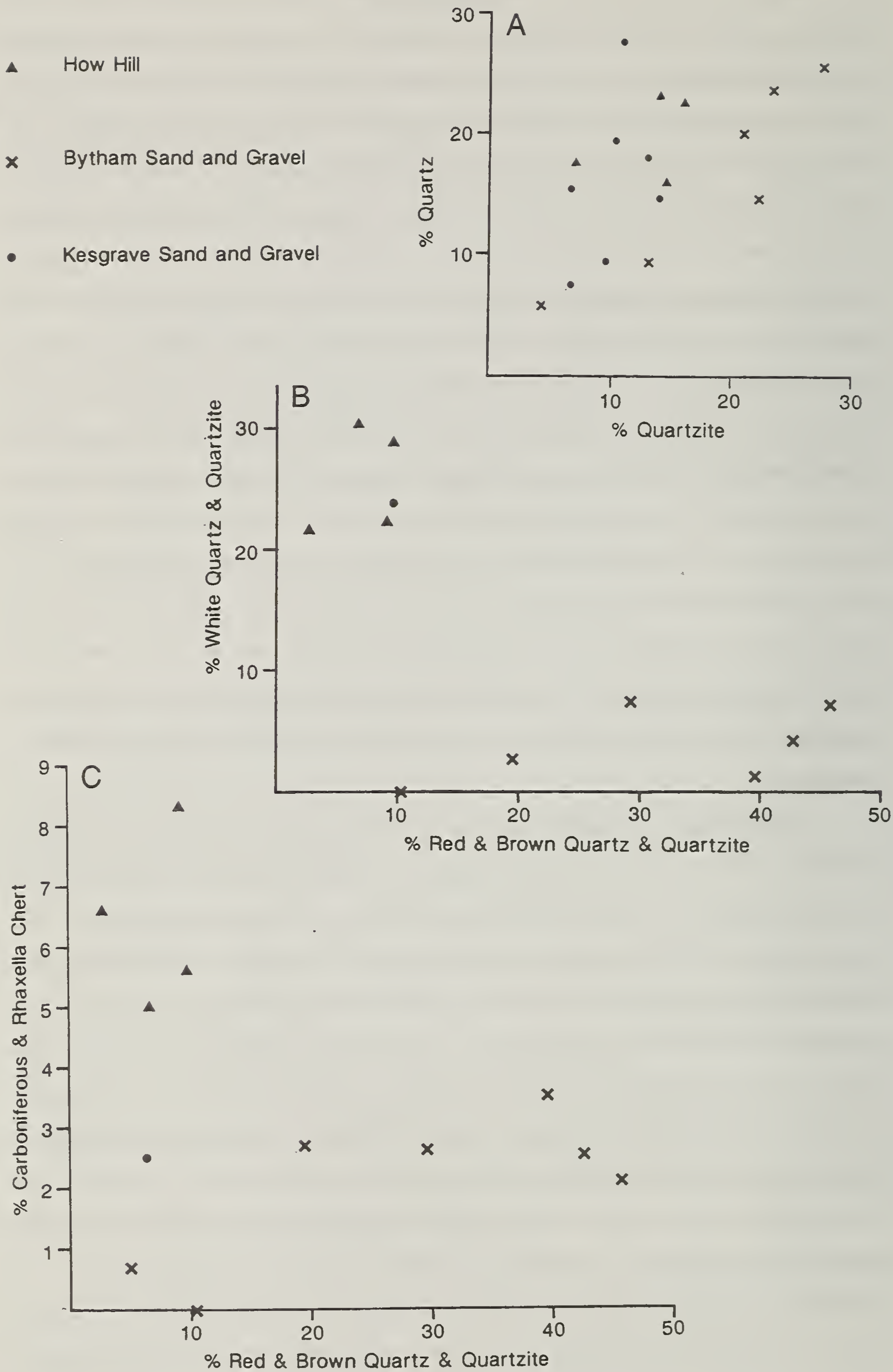
How Hill
Bytham Sand and Gravel
Kesgrave Sand and Gravel
A
% Quartz
% Quartzite
10
20
30
B
% White Quartz & Quartzite
% Red & Brown Quartz & Quartzite
10
20
30
40
50
C
% Carboniferous & Rhaxella Chert
% Red & Brown Quartz & Quartzite
1
2
3
4
5
6
7
8
9

Fig. 8. (facing page)

Graphical representation of the lithological properties of the sands and gravels from How Hill with similar properties from the Kesgrave Sands and Gravels (unpublished and Whiteman and Rose, 1992) and Bytham Sands and Gravels (unpublished). (8A) shows the affinity between samples, according to the quartzite: quartz ratios in the How Hill gravels with the same ratios in the same fraction from the Bytham Sands and Gravels and the 16-32 mm fraction from the Kesgrave Sands and Gravels. This demonstrates the affinity of the How Hill gravels with the Kesgrave Sands and Gravels. (8B) shows the relationship between coloured and colourless quartz and quartzites in the same three lithostratigraphic units. In this case the one sample of Kesgrave Sands and Gravels is represented at the 8-16 mm size range. This indicates the affinity of the How Hill gravels with the Kesgrave Sands and Gravels and the clear separation from the Bytham Sands and Gravels. (8C) shows the relationship between the percentage Carboniferous + Rhaxella chert and the percentage quartz and quartzite, all from the same size fraction. In this case the distribution clearly shows the separation of the How Hill gravels from both the Kesgrave and Bytham Sands and Gravels, indicating the high 'Northern rivers' component.

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derived predominantly from both the Thames and the 'Northern rivers'. This does not rule out a small proportion of Bytham Sands and Gravels that have been reduced in significance by dilution with the other materials.

Assuming a coastal origin, the most likely source of these sands and gravels is coarse-grained material that was transported to the region by rivers with a provenance indicated above.

The diamicton has a clast lithology that reflects entrainment of the common rock types in the underlying sands and gravels with additional flint derived from chalk bedrock to the north.

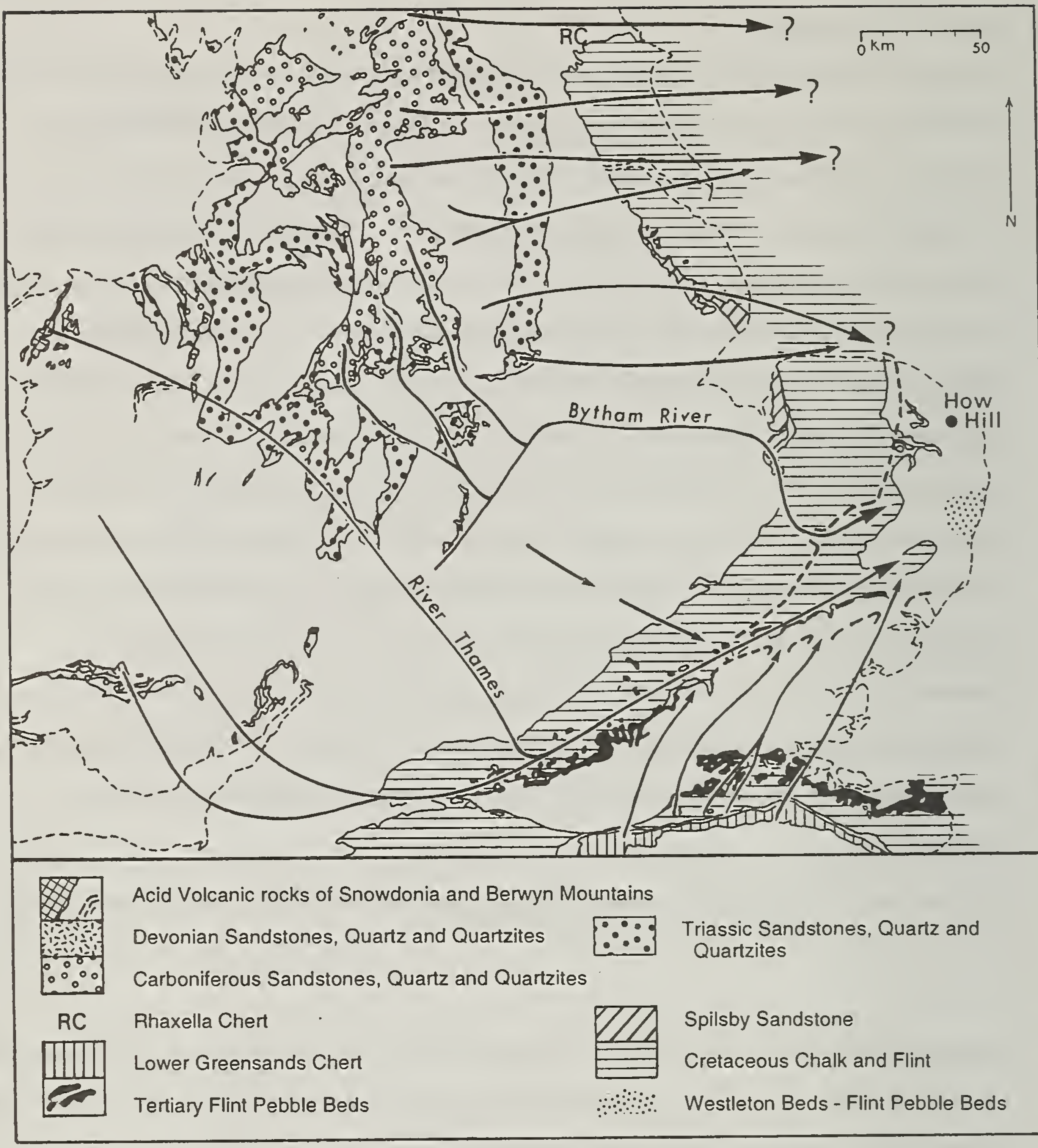
RC
?
?
?
?
0 Km 50
N
How
Hill
Bytham River
River Thames
Acid Volcanic rocks of Snowdonia and Berwyn Mountains
Devonian Sandstones, Quartz and Quartzites
Carboniferous Sandstones, Quartz and Quartzites
Triassic Sandstones, Quartz and Quartzites
RC
Rhaxella Chert
Lower Greensands Chert
Tertiary Flint Pebble Beds
Spilsby Sandstone
Cretaceous Chalk and Flint
Westleton Beds - Flint Pebble Beds

Fig. 9. (facing page)

A tentative outline of the drainage system that could have contributed to the formation of the sands and gravels exposed at How Hill, near Ludham, Norfolk, along with the present outcrop of the main lithologies that may have contributed to the deposits in north east Norfolk. The two main river systems: River Thames and Bytham River are named. The alignment of the rivers is derived from the position of long-established discordant valleys in upland regions, cols of similar origin in hill ranges of eastern England, and the position of river sediments along bedrock valleys (often buried by glacial deposits) in lowland regions and references are given at the appropriate place in the text. This landscape is 'pre-glacial' in age with respect to lowland Britain although glaciers may have existed in the uplands of Wales. The thick dashed line across East Anglia gives the approximate extent of the sediments associated with these rivers in that region, but there is no attempt to differentiate between those deposits that were laid down directly by rivers on a floodplain, braidplain or delta and those deposits which entered the coastal environment and were reworked by waves and tidal currents. It would appear, however that How Hill is within the coastal domain.

DISCUSSION AND WIDER SIGNIFICANCE

The How Hill sections are in a part of East Anglia where natural sections are rare beyond the coastal exposures (West, 1980), and little work has been done on the Pleistocene sands and gravels that exist below the diamicton. Analyses suggest that the lower sorted unit formed as coastal sands and gravels. The main reasons for this interpretation are the sediment types, sedimentary structures and size distribution of the sands which indicate bi-directional flow, variable flow regimes and occasionally high suspended sediment, all typical of tidal current activity. Palaeocurrent directions within the sands indicate that the currents flowed NNW-SSE and SSE-NNW. Tidal current patterns and trends are typically determined by the configuration of the coastline and adjacent shoreline (Stride, 1982; Pearson et al. 1990), but may be either normal or parallel with these constraints as

can be observed in north Norfolk at the present time (Pearson et al. 1990). In these circumstances it seems reasonable to assume that the shoreline was located relatively close to the site, but it is impossible from the sample available to determine the regional shoreline trend. The change from sands to sands and gravels is likely to reflect changes in flow regime associated with shifts in the position of tidal currents as sedimentation progressed. This pattern may be placed in the early Pleistocene palaeogeography shown in Funnell (1991).

The lithologies of the sands and gravels give an indication of the provenance and suggest a source over much of England and Wales. This implies that the region around eastern Norfolk was the depositional focus for the major rivers in existence at this time, a fact that is supported by independent evidence from the North Sea (Cameron et al., 1992, Fig. 104). In this case it would appear that sediment was being contributed by the Thames drainage system from the south with material derived from areas as far afield as Wales, the Midlands and Kent; and by the 'Northern rivers' which drained the southern Pennines, and south Lincolnshire. It seems unlikely that the Bytham river contributed significant amounts of sediment to the deposits at How Hill as its load was diluted by the Kesgrave Sands and Gravels.

There are no lithologies within the sands and gravels that indicate a direct glacial source. For instance, clasts of shale, coal, or metamorphic or igneous erratics from northern England, Scotland or Scandinavia (none of which are destroyed by decalcification), such as are found in the glaciofluvial Barham Sands and Gravels elsewhere in East Anglia (Rose and Allen, 1977) are absent. This negative evidence, combined with the positive affinity with the Kesgrave Sands and Gravels, is the main reason for asserting that the sands and gravels at How Hill are pre-glacial in age. Assuming these correlations to be correct then the deposits at How Hill reflect a very critical time in the Quaternary history of the British Isles in that they show the co-existence of the marine Crag sediments (Mathers and Zalasiewicz, 1988) which are characteristically dominated by local lithologies (Hey, 1976), and the fluviatile Thames and 'Northern rivers' sediments which brought to the region for the first time a major flux of the 'far travelled' lithologies (Whiteman and Rose, 1992). This influx of far-travelled coarse-grained material reflects two factors: i) the time taken for the rivers to transfer these materials down their catchments from the source regions in the uplands of Wales

and south and central England - this may take a 'relatively long time' because the period was initially characterised by low-energy geomorphic systems (Rose, 1988, Whiteman and Rose, 1992); ii) the change in energetics within the upper parts of the drainage basins, associated with a change in the rate and magnitude of orbitally driven climatic forces. Specifically, this appears to have caused glaciation and periglaciation in the mountains of western and possibly central Britain.

The time when this influx of far-travelled lithologies entered the river sediments of eastern England is discussed in detail in Gibbard et al. (1991) and Whiteman and Rose (1992) and is considered to have occurred in the Early Pleistocene in the later part of Tiglian C4c substage of the Netherlands, which is c. 1.75Ma BP (Gibbard et al., 1991; Funnell, 1995).

The diamicton is the decalcified facies of the Cromer Till that was deposited by Scandinavian ice during the Anglian glaciation. The exceptionally well developed wedge within the sands and gravels is interpreted as a water-escape vent with associated collapse and the presence of diamicton within the upper part of this structure indicates that it formed after the till had been deposited. The absence of any evidence for glaciotectonic shear at the contact between the wedge and the till suggests that dewatering occurred in response to the decay of permafrost caused by the insulation of the site by the Anglian ice sheet or by climatic amelioration at the end of the Anglian glacial stage.

ACKNOWLEDGEMENTS

This work was funded by a BGS Research Grant, GA/94A/23. Sincere thanks are expressed to Drs. C.A. Whiteman, P. Allen, C.P. Green and R.W. Hey and R.A. Kemp, and Professor B.M. Funnell, for very helpful critical comments on the first draft of the manuscript. Constructive comments by the referees, Drs. G. Postma and Dr. P.L. Gibbard are also acknowledged. Dr. C.P. Green is also thanked for advice with the identification of clasts and Mr. J. Jacyno for help with the cartography. We would also like to thank Mr. P. Boardman who permitted access to the land on which the trial pits were excavated. With respect to B.S.P.M., R.J.O., D.H.J and J.B.R, this paper is published with permission of the Director, British Geological Survey (N.E.R.C.).

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CARBON AND SULPHUR GEOCHEMISTRY AND CLAY MINERALOGY OF THE WEST RUNTON FRESHWATER BED

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ABSTRACT

Weight % organic carbon and weight % total sulphur (C/S) in the Cromerian West Runton Freshwater Bed (WRFWB) of north Norfolk are not simply related to depositional environment as in the classic Berner and Raiswell (1984) model. This is because the sediment is organic matter-rich, with total sulphur representing a variable mixture of pyrite sulphur and organic sulphur. In addition, remobilisation of sulphur associated with post-depositional groundwater flow has modified depositional values. In the upper 20 cm of the bed, pyrite sulphur has been oxidised and largely removed. This sulphur was subsequently reprecipitated as later diagenetic pyrite at about 40 - 60 cm in the bed under reducing conditions. Below 60 cm C/S ratios are probably close to depositional values and can be used as palaeosalinity indicators. As expected, C/pyrite-S ratios in this lower part of the bed indicate a freshwater depositional environment.

The groundwater which oxidised pyrite in the top of the bed also destroyed organic carbon and probably dissolved aragonitic shell material. However, preservation of aragonitic shells below 60cm in the WRFWB defines the extent to which groundwater penetrated the bed.

Smectite is one of the dominant clay minerals in the WRFWB and potassium saturation suggests that the smectite may have had a volcanic origin. The Eifel region of Germany is a possible source of fine-grained ash or dust at this time.

Bull. geol. Soc. Norfolk (for 1995) **45**, 29-51 (Published 1996)

INTRODUCTION

The Cromerian sedimentary rocks which crop out on the Norfolk coast between Sheringham and Cromer (West 1968; 1980), are very well known to geologists of the Quaternary, because of their international importance as the type site for the Cromerian interglacial period of the early Middle-Pleistocene. The West Runton Freshwater Bed (hereafter abbreviated to *WRFWB*, equivalent to the whole of bed f in West 1980, p.17), part of the Cromer Forest-bed Formation (Funnell and West 1977; Gibbard and Zalasiewicz 1988; Gibbard et al. 1991), though a localised deposit, is similarly famous because of its well-preserved terrestrial and freshwater fossils, not least because of the recent discovery and successful excavation of a large and well-preserved elephant skeleton (Stuart 1991). There has, however, been little or no detailed work on the clay mineralogy and geochemistry of the bed.

In this paper we combine carbon and sulphur geochemistry with aspects of the sedimentology and mineralogy of the bed in an attempt to constrain further the environment of deposition, and more important, to suggest how post-depositional alteration has affected the deposit. The top of the *WRFWB* (upper 20 cm) is of particular interest in this context, as it has clearly been altered, both in colour and chemical composition (subsequently referred to as the oxidised Fe bed), compared with the main body of the bed. Stuart (1991) - reflecting a widely thought, though unpublished view - suggested the oxidised Fe bed may be a palaeosol (N.B. not the same horizon as that described by Valentine & Dalrymple (1975)), indicating a period when the bed was exposed to subaerial processes. However, West (1980) described this altered layer only as being 'stained red by iron oxide', suggesting some uncertainty as to its origin.

Carbon/sulphur geochemistry: background

It has been demonstrated that organic carbon/pyrite sulphur (C/S) ratios can be used as a palaeosalinity indicator to distinguish marine from freshwater sedimentary rocks (Berner & Raiswell 1983, 1984). The differences in C/S ratios between marine and freshwater environments are caused by the process of sedimentary pyrite formation.

In organic-rich marine sediments, pyrite formation is a ubiquitous process under anaerobic conditions. Sulphate (SO_4^{2-}) present in the porewater of the anoxic sediment provides an electron acceptor for the bacterial oxidation of organic matter. Simultaneous reduction of the sulphate to hydrogen sulphide (H_2S) also occurs.

$$2CH_2O + SO_4^{2-} \rightarrow H_2S + 2HCO_3^- \qquad [1]$$

The H_2S is available to react with detrital iron minerals (usually present as oxidised grain coatings), to form iron monosulphides (FeS). These monosulphides subsequently react with sulphur to produce pyrite (FeS_2) during early diagenesis.

$$3H_2S + 2FeOOH \rightarrow 2FeS + S + 4H_2O \qquad [2]$$

$$FeS + S \rightarrow FeS_2 \qquad [3]$$

In anoxic freshwater sediments little pyrite is formed due to low concentrations of SO_4^{2-} The average concentration of dissolved SO_4^{2-} in seawater is 28 mmol l^{-1}, whereas freshwater SO_4^{2-} concentration is approximately 1% of this value (Berner & Raiswell 1984).

The C/S ratios in modern normal marine sediments (conditions of oxygenated rather than anoxic bottom water, Raiswell & Berner 1986), averages at 2.8. In freshwater sediments the C/S ratio is much higher, typically values of 15 or more, because less pyrite sulphur is present (Berner & Raiswell 1984). Hence, in freshwater environments, pyrite formation is limited by the low concentration of SO_4^{2-}, whereas in marine environments reactive organic matter (i.e. that which can be assimilated by bacteria), is the principal limiting factor. Aplin and Macquaker (1993) have suggested recently that pyrite formation in marine sediments is also controlled by the rate at which H_2S reacts with iron phases.

Plotting weight % organic carbon against weight % total sulphur, differentiates freshwater and marine sediments (Fig. 1). The linear relationship for modern marine sediments is represented by the 'marine line', but no strong positive correlation between organic carbon and pyrite sulphur exists in freshwater sediments. In this respect, the

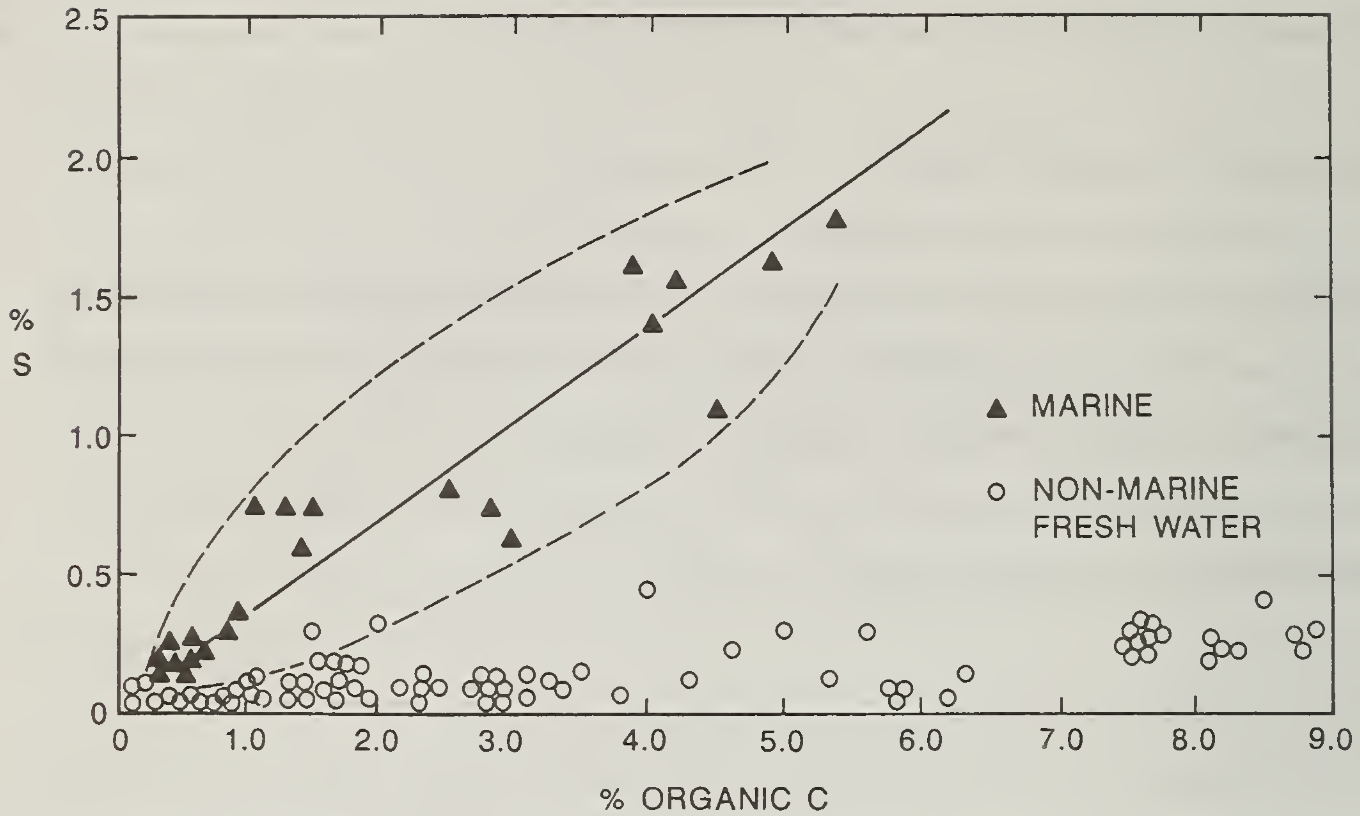

Fig. 1. Plot of wt% organic carbon and wt% pyrite sulphur for modern freshwater lake sediments and normal marine sediments (after Berner & Raiswell 1983).

WRFWB appears to be an interesting target for carbon/sulphur geochemistry since the palaeosalinity of the bed is well constrained by palaeontological data (Sparks 1980, Stuart 1975, 1991, West 1980).

There are, however, limitations to the applicability of the C/S palaeosalinity method. These limitations include the use of:

(1) sediments and rocks with <1 wt% C, such as sandstones, where palaeosalinity is impossible to distinguish due to low sulphur concentrations in both marine and freshwater situations;

(2) sediments and rocks with >15 wt% C, such as coal, which have high organic sulphur contents that consequently increase the total sulphur content (Altschuler *et al.* 1983). If total sulphur cannot be reasonably assumed to represent pyrite-S, then pyrite-S must be determined specifically. For example, Davison *et al.* (1985), in a study of modern freshwater lake sediments concluded that total sulphur was not a good approximation of pyrite sulphur in their non-marine sediments;

(3) sediments and rocks with a negligible iron content which would result in iron limitation during pyrite formation, leading to a low FeS_2 content and consequently a high C/S ratio. This eliminates most carbonate sediments and rocks. Saltmarsh sediments are also known to exhibit iron limited pyrite formation, even though SO_4^{2-} concentrations are high. Here the rate of pyrite formation is determined by the sulphidation of refractory (rather than reactive) iron (Lord & Church 1983);

(4) highly weathered samples where oxidation alters the C/S ratio by reducing both the organic carbon and pyrite sulphur content of the sediment.

Limitations 2 and 4 might be expected to apply to the *WRFWB* which is quite organic matter rich, and certainly weathered in its upper part (oxidised Fe bed).

Site Description and the Palaeoenvironment at West Runton

The *WRFWB* is exposed at the base of coastal cliffs at West Runton, north Norfolk. The best exposures are found to the east of Woman Hythe (West Runton Gap), between national grid references TG 186431 and TG 189431 (Fig. 2).

The *WRFWB* consists mainly of organic-rich detrital mud (bed f) which becomes progressively fossiliferous toward the base (Fig. 3). Between the Upper Cretaceous Chalk on the modern wave cut platform and the *WRFWB*, shelly marine sands, gravels and silts of Pre-Pastonian, Pastonian and Beestonian age are intermittently exposed (parts of the Runton, Paston and Sheringham Members of West (1980, tables 8 and 43)). The temperate stage deposits of bed f are overlain by marine transgressive tidal sands (Mundesley Member, West (1980, table 43)), and a final regression to freshwater deposits (Bacton Member, West (1980, table 43)) as climate cooled before the onset of the Anglian

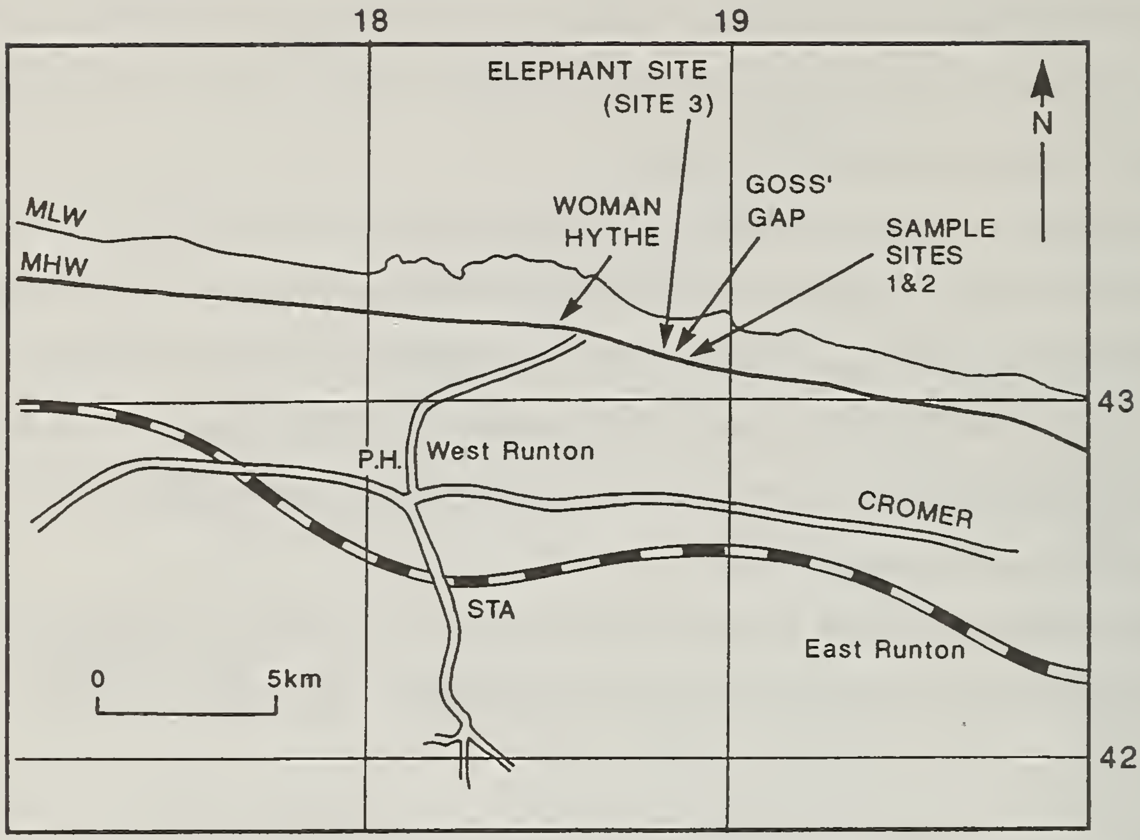

Fig. 2. Location map of samples sites at West Runton (Ordnance Survey grid lines). Sample sites 1 and 2 are close to section WRCGC of West (1980, fig.6) where bed f is about 90cm thick.

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glaciation (Jones & Keen 1993). The Cromer Till (part of the North Sea Drift) represents the first ice advance of the Anglian stage glaciation (Fig. 3).

The *WRFWB* probably represents a freshwater channel, infilled during the temperate Cromerian stage (West 1980). Palynology and palaeontology have given a detailed picture of the depositional facies. Pollen assemblages indicate a regional mixed oak woodland of thermophilous trees (*Quercus* (oak), *Ulmus* (elm), and *Tilia* (lime)), assigned by West (1980) to the CrIIb pollen substage. These trees probably flanked the river valley, with fen and herbaceous vegetation occupying the floodplain (Stuart 1975). The *WRFWB* has yielded a rich vertebrate fauna, which includes 43 mammals, 17 birds, 8 fish taxa, 5 amphibians and 3 reptiles (Stuart 1991). The discovery of a large elephant

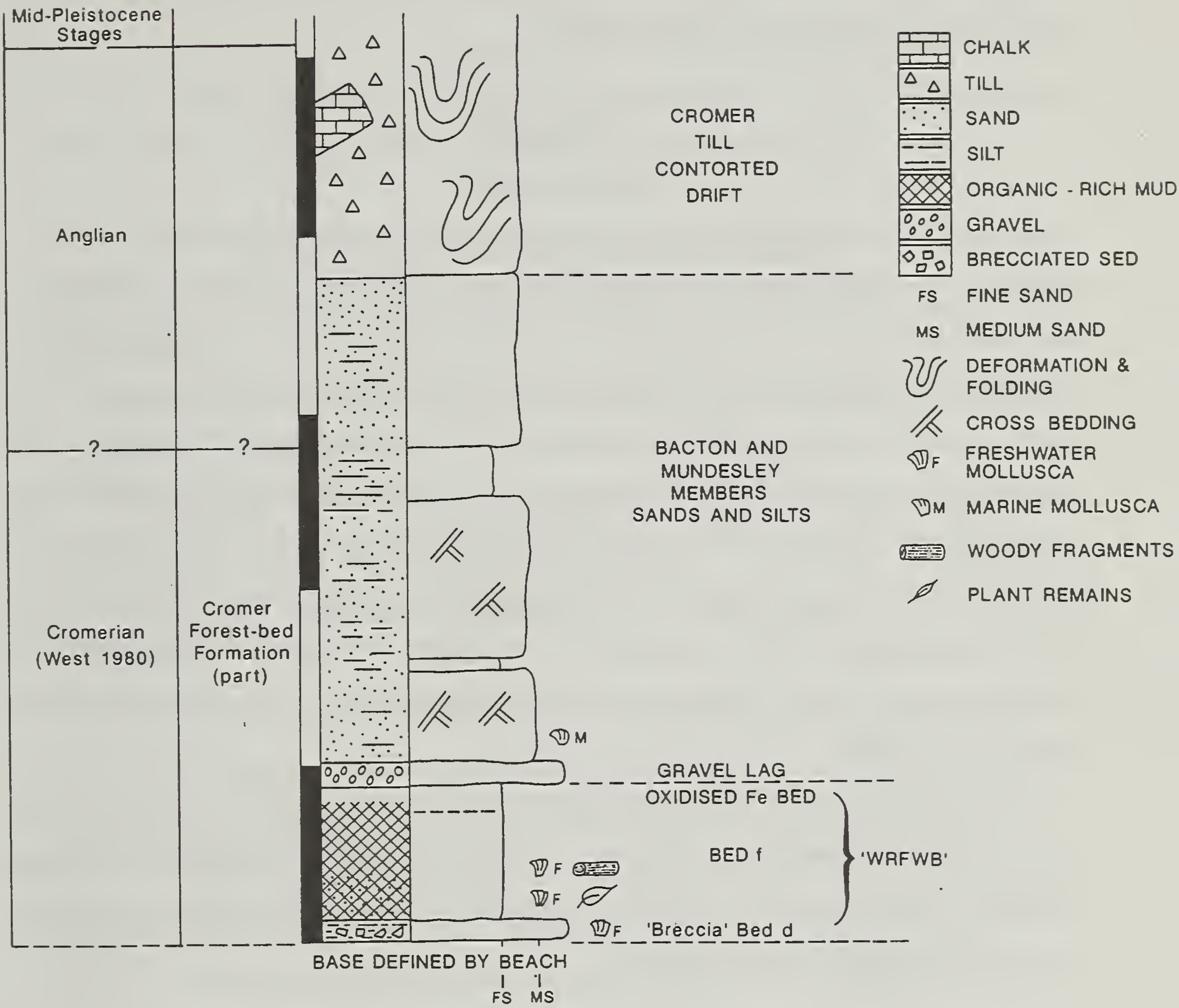

Fig. 3.

Summary stratigraphy and sedimentary log of the cliff section at West Runton, east of Goss' Gap, as exposed in June 1994. Black and white scale bars = 1m. Beds d and f of the *WRFWB* after West (1980). Bed f comprises the main part of the *WRFWB* (= West Runton Member of Funnell and West (1977)). Mundesley and Bacton Members of West (1980, Table 43)

(*Mammuthus trogontherii)* pelvis, in 1990, led to the excavation of most of the skeleton (excavation completed in December 1995). Freshwater molluscs within the bed f indicate a slow moving, well vegetated body of water (Sparks 1980), while the land mollusca are characteristic of marshy, water edge conditions.

METHODS

Sample collection

Two sites a few metres east of Goss' Gap (Fig. 2) - close to section WRCGC of West (1980, fig. 6) - were selected for sample collection in June 1994. At site 1, 11 samples were taken at 10 cm intervals through bed f from the oxidised Fe layer to the base of bed f. The first three samples were taken from the oxidised Fe layer at about equal intervals. The weathered surface of the sediment was scraped away to obtain a fresh sample, typically 10-20g weight. Sandy and $CaCO_3$-rich areas of the bed were avoided where possible to comply with the limitations of the C/S ratio method (see above). The process was repeated for site 2, although only one sample was obtained from the oxidised Fe bed. Bulk samples for clay mineralogy (50-100 g of sediment), were obtained from the middle of the oxidised Fe bed and bed f at site 2. Shell fragments from the lower part of bed f were collected to determine mineralogy.

In November 1995, a third section at the West Runton Elephant excavation (Fig. 2) -- a particularly clean and unweathered site -- was sampled for determination of sulphur speciation. These samples were freeze dried within three hours of collection to minimise oxidation of metastable sulphur species.

Carbon and sulphur chemistry

The total carbon and sulphur content of samples from sites 1 and 2 were measured using a Carlo Erba EA 1108 elemental analyser (UEA). Total sulphur analyses for sulphur speciation (site 3) were measured using a Carlo Erba EA 1106 elemental analyser (Leeds). Both machines were calibrated with a sulphanilamide ($C_6H_8N_2O_2S$) standard. Acidified chromous chloride was used to extract and determine inorganic sulphide-S, (probably pyrite since no acid volatile sulphur was produced upon initial acidification). This process removes both sulphidic-S and acid soluble sulphatic-S, but has minimal effect on organic

bound S (Bottrell et al. 1994). Organic-S was determined by Carlo Erba elemental analysis of the dried chromous chloride extraction and corrected for weight loss.

Organic carbon content was determined by reacting 1-2 g of powdered sample with 10% hydrochloric acid for one hour to remove $CaCO_3$. The sediment was then filtered and washed thoroughly with distilled water and dried at 80°C overnight before elemental analysis. Four blanks and five standards were run before samples.

Clay mineralogy and sediment micromorphology

Clay minerals were identified on pre-treated samples by X-ray diffraction (XRD). Pre-treatment included:

a) reaction with hydrogen peroxide to oxidise organic matter and release clay minerals bound to organic matter;
b) sieving through a 63 μm mesh to reduce the sample to less than coarse silt size;
c) reaction with sodium acetate to remove carbonate;
d) saturation with magnesium chloride to improve XRD peak definition.

The final stage was to isolate the <2μm fraction by the pipette method (Brimblecombe *et al.* 1982). Three aliquots of the <2μm fraction were produced as smear slides for XRD and clay minerals were identified on the basis of their air dried, glycolated and heat treated diffraction peak positions (after Biscaye 1965). In addition, the response of smectite clay to potassium saturation (K-saturation - as outlined by Weaver (1958)) was investigated to provide information on the nature of the parent material from which the clay formed.

A small block of undisturbed sample from the oxidised Fe bed was collected by inserting a plastic film case into the sediment. The sample block was dried, resin impregnated and thin-sectioned for optical microscopy.

RESULTS

Carbon and sulphur

Carbon and sulphur data from sites 1 and 2 (Table 1) are plotted on Fig. 4. High carbon values were expected due to the organic-rich nature of the bed. Surprisingly there were also high sulphur values, an unexpected result for freshwater sediments where low concentrations of SO_4^{2-} usually limit pyrite formation (Berner and Raiswell 1984).

The four samples containing no sulphur (plotting on the X-axis of Fig. 4) are from the oxidised Fe bed. The oxidised Fe bed has low carbon and sulphur values, statistically different (t-test) from values for the rest of bed f. However, sulphur values are variable within bed f: at approximately 11 wt%C, the wt% sulphur values range from 1.9 to 4.2.

At both sites, C/S ratios in the oxidised Fe bed are highest (values of 6-7; Fig. 5) decreasing (stratigraphically) down the bed and becoming more constant below 40 cm depth (to values around 2.5; Fig. 5). Taken at face value, these ratios overlap those of normal marine sediments (Fig. 4) since freshwater C/S ratios are expected to be 15 or more.

The sulphur speciation data from site 3 (Table 2) confirm that the oxidised Fe bed (10 and 20cm samples) contains very little sulphur, while the sample below (30cm) has only 0.56wt% S, and is the only sample to contain sulphatic-S (0.1wt%), i.e. its total S is significantly in excess of sulphidic + organic sulphur. The next sample below (40cm) has the highest sulphur content (2.04wt% organic-S and 3.82wt% pyritic-S). Below this, organic-S concentrations are variable but low (<0.93wt%) while sulphide-S concentrations fall to progressively lower values (minimum of 0.47wt%) toward the base of the bed.

X-ray diffraction data

Clay Mineralogy

Smectite and vermiculite were identified as the dominant clay minerals in both bed f and the oxidised Fe bed. The broad peak between 14.6 and 15.0Å on the air dried XRD trace, mainly expanded to 17Å on glycolation and collapsed to 10Å on heating to 440°C. As smectite and vermiculite peaks overlap it is unclear if these minerals were present as discrete phases or as an interstratification. Response to glycolation suggests smectite is the dominant clay in the mixture and it is possible that the vermiculite is a modern weathering

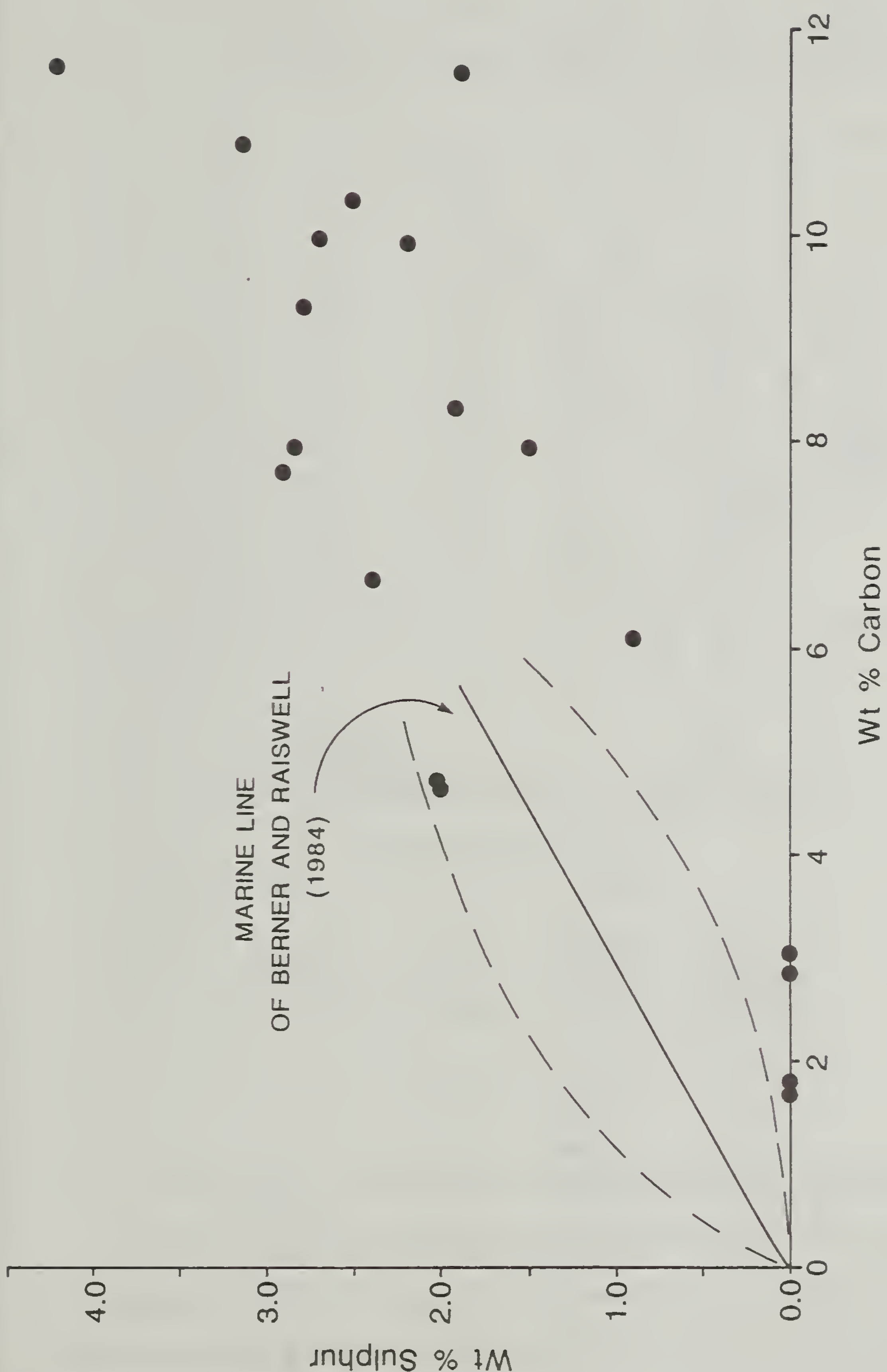

Fig. 4. Wt% organic carbon and wt% total sulphur for the *WRFWB*. Standard errors are contained within data points.

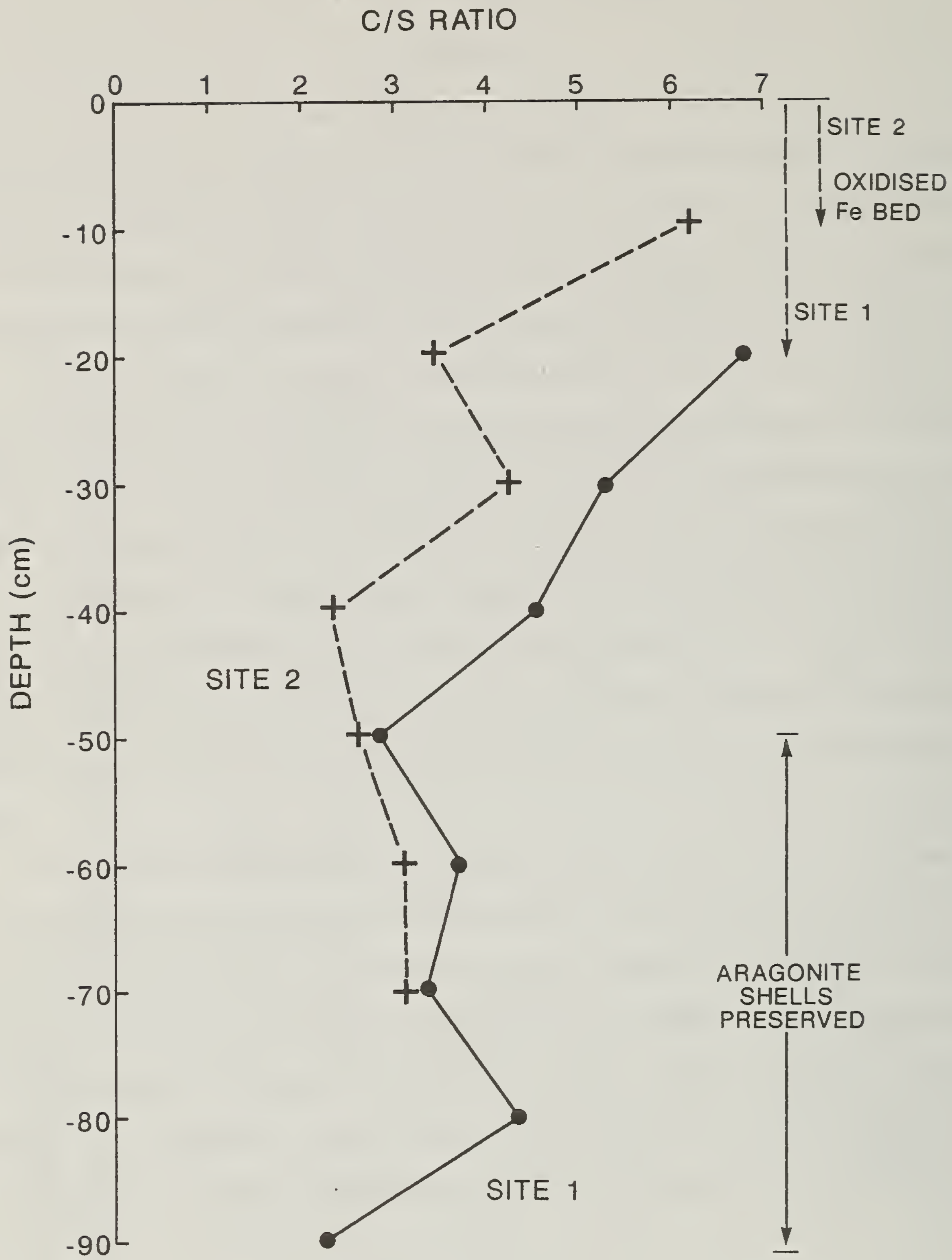

Fig. 5.

C/S ratios plotted against depth in the f bed at sites 1 and 2 (where 0 = the top of the *WRFWB*). The thickness of the oxidised Fe bed, and the presence of aragonitic shells is also indicated. Note that the lack of aragonitic shells in the upper 50cm of bed f coincides with the change from C/total-S ratios around 3 (below 50cm) to a gradual increase toward values of 6 or 7 (above 50cm). This is consistent with the upper 50cm of the bed being weathered, causing the oxidation of pyrite and dissolution of aragonite.

Table 1. Carbon and sulphur data - West Runton Freshwater Bed

Depth below top of bed (cm)	Wt % Carbon (Site 1)	Wt % Sulphur (Site 1)	C/S ratio (Site 1)	Wt % Carbon (Site 2)	Wt % Sulphur (Site 2)	C/S ratio (Site 2)
3	-	-	-	3.05	0	-
4	1.69	0	-	-	-	-
9	1.80	0	-	-	-	-
10	-	-	-	11.57	1.89	6.12
13	2.86	0	-	-	-	-
20	6.10	0.90	6.78	10.88	3.14	3.46
30	7.94	1.50	5.29	10.33	2.51	4.12
40	9.92	2.19	4.53	4.67	2.01	2.32
50	6.66	2.39	2.79	7.71	2.91	2.65
60	9.96	2.70	3.69	7.95	2.84	2.80
70	9.20	2.79	3.33	11.63	4.21	2.76
80	8.32	1.92	4.33	-	-	-
90	4.71	2.02	2.33	-	-	-

Table 2. Sulphur speciation data - West Runton Freshwater Bed. Site 3.

Depth below top of bed (cm)	Wt % Organic Sulphur	Wt % Pyrite Sulphur	Wt % Sulphatic Sulphur
10	<0.02	<0.002	<0.02
20	<0.02	0.003	<0.02
30	0.13	0.33	0.10
40	2.04	3.82	<0.05
60	0.30	1.59	<0.05
80	0.04	0.97	<0.05
100	0.93	0.94	<0.05
120	0.15	0.47	<0.05

product of the smectite (see discussion in Andrews 1987). In bed f the smectite and vermiculite content was 79% and in the oxidised Fe bed they constituted 85% of the clay minerals. Kaolinite was the secondary clay mineral, identified by a 7Å peak that was unaffected by acid treatment, but collapsed when heated to 550°C. Kaolinite constituted 20% and 14% of the clay mineralogy in bed f and the oxidised Fe bed respectively. Chlorite and illite were only present in trace amounts.

K-saturation of smectite caused only a partial collapse of the 14.6 -15.0Å peak and incomplete migration toward 10Å (Fig. 6) suggesting a possible volcanic origin of the smectite (see Weaver 1958).

$CaCO_3$ Shell Mineralogy

X-ray diffraction showed that mollusc shells within bed f were aragonitic with some gypsum also present.

DISCUSSION

C/S ratios in the West Runton Freshwater Bed

The C/total-S ratios from most the *WRFWB* (average 3.82 ± 1.38) generally overlap an extension of the Berner and Raiswell (1984) 'marine line' (average normal marine C/S ratio = 2.8). Since the f bed is clearly a freshwater deposit (see above), the apparently marine C/total-S signature must be an artifact of the high total sulphur content of the sediment. The sulphur speciation data (Table 2) show that most of the non-pyrite sulphur is organic sulphur.

Organic sulphur (carbon bonded sulphur) compounds are derived from organic plant and animal remains, and are usually present as amino acids and proteins. Organic-rich sediments, such as peat and coal, often have an appreciable sulphur content (e.g. 0.1 to 10% in coals (Ruch *et al.* 1974) and organic sulphur can represent more than 50% of the total sulphur in coal (Casagrande & Ng 1979).

The sulphur speciation data (Table 2) reveals important additional information. Below the oxidised Fe bed, the upper part of the f bed (40 - 60cm below the top of the bed) still contains more pyrite-S than is usually found in freshwater sediments (Fig. 1). However, below this level (i.e. more than 60cm below the top of the bed), the lower pyrite-S values are compatible with a freshwater environment.

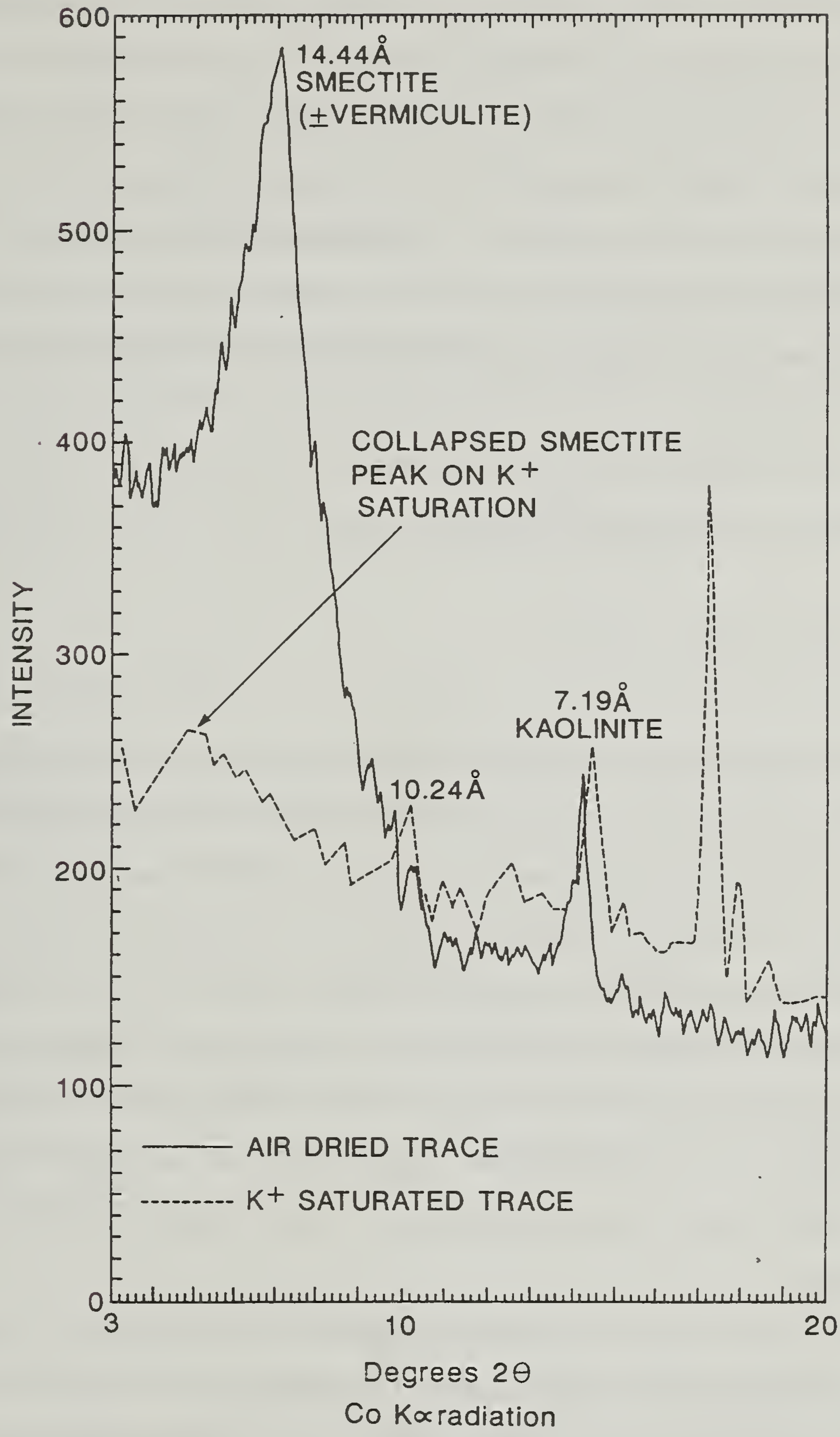

Fig. 6. Smectite XRD trace in solid line. Dotted line represents clay sample that has been K-saturated.

We suspect that this pattern has been produced in part by post-depositional diagenetic remobilisation of sulphur in the bed. The thin section from the oxidised Fe bed contained no clear soil fabric (see Fitzpatrick (1980)) or other soil characteristics such as burrows or evidence of translocated clay. On this basis we do not interpret the oxidised Fe bed as a palaeosol. Instead we interpret it as a weathering horizon, probably caused by post-burial, subsurface flow of groundwater. The characteristic red/rust colour of the bed indicates the presence of oxidised iron and surficial patches of powdery yellow elemental sulphur suggest that sulphur-bearing compounds (i.e. pyrite and organic matter) have been oxidised:

$$2FeS_2 + 2H_2O + 7O_2 \rightarrow 4H^+ + 2Fe^{2+} + 4SO_4^{2-} \qquad [4]$$

$$CH_2O + O_2 \rightarrow CO_2 + H_2O \qquad [5]$$

The oxygen in the groundwater would have been consumed by both the reactions shown above, such that as groundwater translocated soluble SO_4^{2-} into the still reduced lower part of the bed below, sulphate reducing bacteria were able to use the remobilised SO_4^{2-} as an electron acceptor, resulting in the formation of more (i.e. later diagenetic) pyrite (reactions 1-3). This explains the anomalously high pyrite sulphur values at 40 - 60 cm depth in the bed and we are currently testing this hypothesis using sulphur isotope geochemistry.

If our explanation above is correct, this means that only the lower samples (below 60 cm in the bed) carry a geochemical signal related to the original depositional environment. These samples have lower pyrite-S values, consistent with a freshwater palaeoenvironment.

The low values of organic carbon in the oxidised Fe bed are also neatly explained as a consequence of this oxidation. Moreover, the CO_2 produced by reaction 5, could have contributed to the acid hydrolysis of $CaCO_3$, which is indicated by the absence of shells in the oxidised Fe bed and the upper part of bed f. Sparks (1980) also attributes the lack of shelly material at the top of bed f to weathering rather than an original absence of shells. The presence of aragonitic shells in the basal 40cm of bed f is good evidence that dissolution by groundwater has not occurred at this level and defines the extent of the weathering horizon.

Total oxidation of pyrite in the oxidised Fe bed would have yielded iron oxides and acidic solutions:

$$2FeS_2 + 2H_2O + 7O_2 \rightarrow 4H^+ + 2Fe^{2+} + 4SO_4^{2-} \qquad [6]$$

$$4Fe^{2+} + 2O_2 + 10H_2O \rightarrow 4Fe(OH)_3 + 8H^+ \qquad [7]$$

The resulting iron hydroxide ($Fe(OH)_3$), would have dehydrated to form an iron oxide, giving the oxidised Fe bed its characteristic colour. The acidic solution would not only have hydrolysėd $CaCO_3$, but the resulting ions from both reactions (SO_4^{2-} from iron sulphide oxidation and Ca^{2+} from $CaCO_3$ acid hydrolysis) could have combined to form gypsum,

$$CaCO_3 + H^+ \rightarrow Ca^{2+} + HCO_3^{2-} \qquad [8]$$

$$Ca^{2+} + SO_4^{2-} + 2H_2O \rightarrow \underset{\text{(gypsum)}}{CaSO_4.2H_2O} \qquad [9]$$

which was found associated with the shell fragments during XRD analysis. Complete $CaCO_3$ dissolution in the upper part of the bed would explain the presence of gypsum lower in the bed. Sparks (1980) described well-preserved gypsum crystals at the top of bed f, and concluded, like us, that they are a diagenetic product of combined pyrite oxidation and carbonate dissolution.

Clay mineralogy

Smectite was identified as the dominant clay mineral in both bed f and the oxidised Fe bed. Smectite has two major sources: (i) crystallisation from soil solutions weathering minerals such as plagioclase feldspar, biotite and muscovite (the latter provide the source of Fe and Mg in smectite); and (ii) from the weathering of volcanic glass (Berner 1971). The K-saturation test indicated a possible volcanic origin for the smectite in the *WRFWB*.

Although no proximal volcanic source existed at West Runton during the Cromerian, ash clouds could have travelled from relatively distant eruptions. During the Cromerian, volcanism was occurring in the Eifel province of Germany. The Wehr volcano is one of the three major centres in the East Eifel volcanic field (near Bonn), and has been active over the past 500,000 years (Wörner *et al.* 1988). These eruptions may have provided the volcanic material (Na, K, Al, and Si-rich glassy minerals) for the source of smectite found in the *WRFWB*. This is a tentative suggestion, although subsequent Quaternary eruptions of the Laacher See volcano in the East Eifel deposited material in northern Italy and southern Sweden (Büchel *et al.* 1986; see also Fig. 7) showing that fine-grained ash is capable of travelling long distances.

A soil-zone weathering source for the formation of smectite cannot be dismissed. Although the 14.6-15.0Å XRD peak only partially migrated during K saturation, it was still relatively close to the 10 Å position (Fig. 6). Mineralogical composition of the source rock is not the only factor which determines the type of clay mineral formed; relief and rate of fluid flow are also important. If water flow rates are sluggish, the reaction of cations with $Al(OH)_3$ and Si are also slow, resulting in the formation of smectite (Berner 1971). Hence, when flushing rates are low, only the more reactive rock components (especially volcanic glass) are decomposed. Relief also retards leaching because topographic depressions are often poorly drained, allowing negligible water flow (Raiswell *et al.* 1980). Smectite would be formed in environments that exhibit these characteristics, such as swampy lowlands. The palynological and palaeontological evidence from the *WRFWB* suggested that these environmental conditions existed at this location, possibly encouraging smectite formation.

In summary, the presence of smectite in the *WRFWB* could be due to the physical characteristics of the depositional environment although volcanic ash may have been a component of the parent material. These preliminary results require further work. In particular we need to know more about the stratigraphic distribution of smectite within the bed. If the smectite is in discrete layers it might represent alteration of contemporaneous air-falls of volcanic dust, whereas a more homogeneous distribution of smectite might suggest the clay was washed into the deposit from older, pre-*WRFWB* sediments.

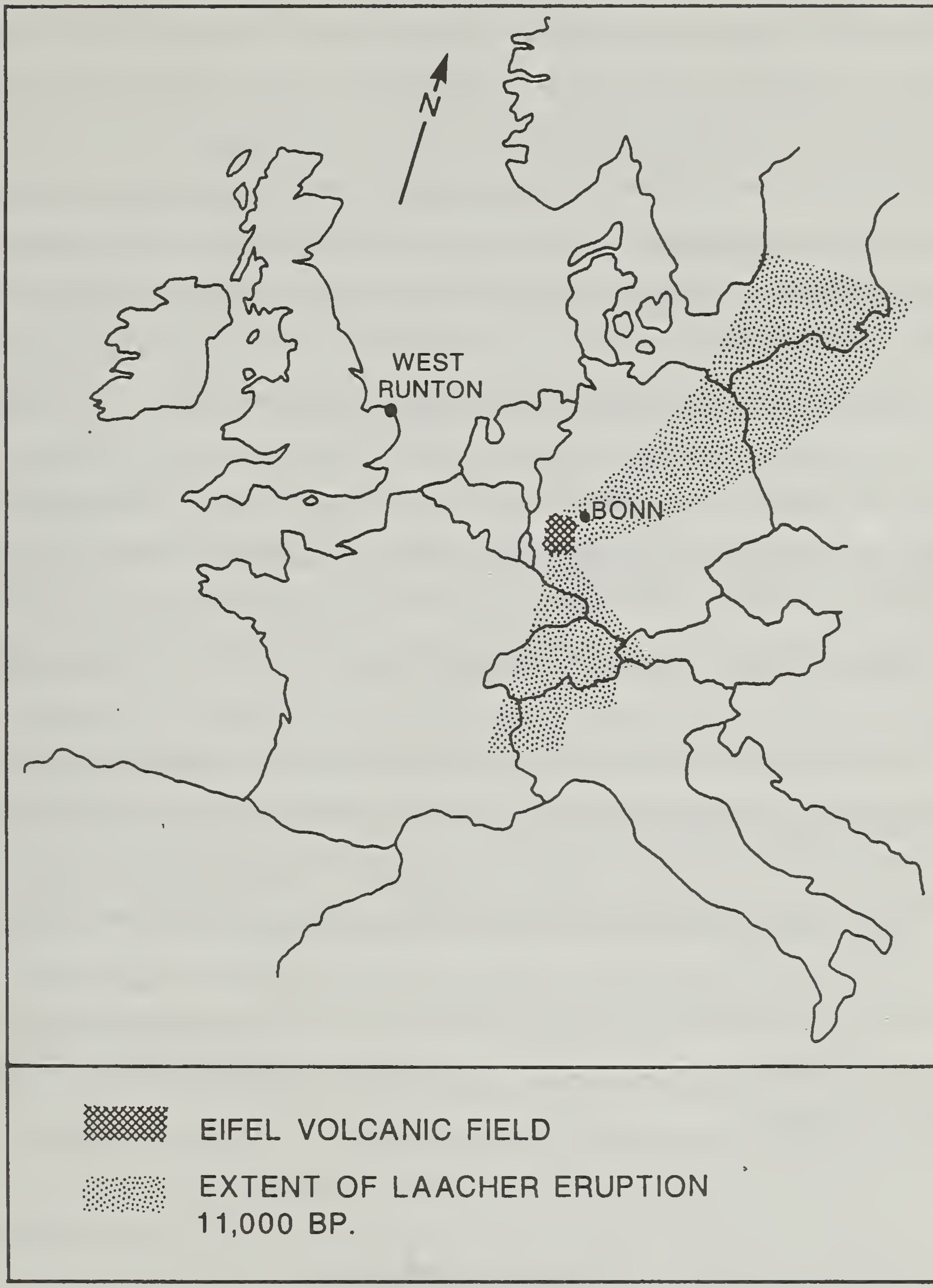

Fig. 7. Location of the Eifel volcanic region, Germany and the extent of the Laacher See eruption. It is not inconceivable that the *WRFWB* could have received volcanic input from this or other similar eruptions.

CONCLUSIONS

1. The carbon and sulphur geochemistry of the West Runton Freshwater Bed is not simply related to depositional environment. Remobilisation of sulphur associated with groundwater flow has changed depositional values. In the uppermost parts of the bed, pyrite sulphur has been oxidised and largely removed. However, this sulphur subsequently reprecipitated as later diagenetic pyrite at about 40-60 cm in the bed. This is an important finding, which shows that diagenetic effects can both decrease and increase pyrite sulphur relative to depositional values. Below 60 cm in the bed, carbon/sulphur ratios are probably close to depositional values and can be used as palaeosalinity indicators.

2. Our data suggest that sediments with <10wt% organic carbon, rather than <15wt% (Berner and Raiswell 1984), might be a better upper threshold value for the C/S palaeosalinity technique. Total sulphur concentrations in the *WRFWB* do not represent pyrite sulphur and they may not do in other similar deposits with >10wt% organic carbon.

3. The top of the West Runton Freshwater Bed has been altered. This oxidised-Fe bed has no characteristic soil structure and, hence, probably represents a weathering horizon related to groundwater flow. The preservation of aragonitic shell fragments towards the base of the *WRFWB* is an indication that groundwater did not penetrated into this part of the bed.

4. Smectite was identified as the primary clay mineral within the bed, and its source may be volcanic. The closest volcanic source to West Runton during the Pleistocene was the Eifel region in Germany.

ACKNOWLEDGMENTS

Thanks are due to Tony Stuart for allowing us to sample at the elephant excavation site. Liz Rix, Stephen Bennett, Ian Marshall, and Rick Bryant (UEA) and Dave Hatfield (Leeds) gave technical assistance in the laboratory and Christine Flack and Philip Judge (UEA) helped with the manuscript and diagrams. We are very grateful to the Rob Raiswell, Brian Funnell and Tony Stuart for their constructive criticisms of earlier versions of this paper.

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CONTENTS

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Copies of the Bulletin may be obtained from the Secretary at the address given above; it is issued free to members.

The illustration on the front cover is figure 4 from the article by Hannam *et al.* in this issue of the Bulletin. It shows the weight percentage of organic carbon and total sulphur in samples of the West Runton Freshwater Bed. Some of the sulphur values are high for a freshwater deposit, and this, along with other evidence, suggests that sulphur has been remobilised some time after deposition.

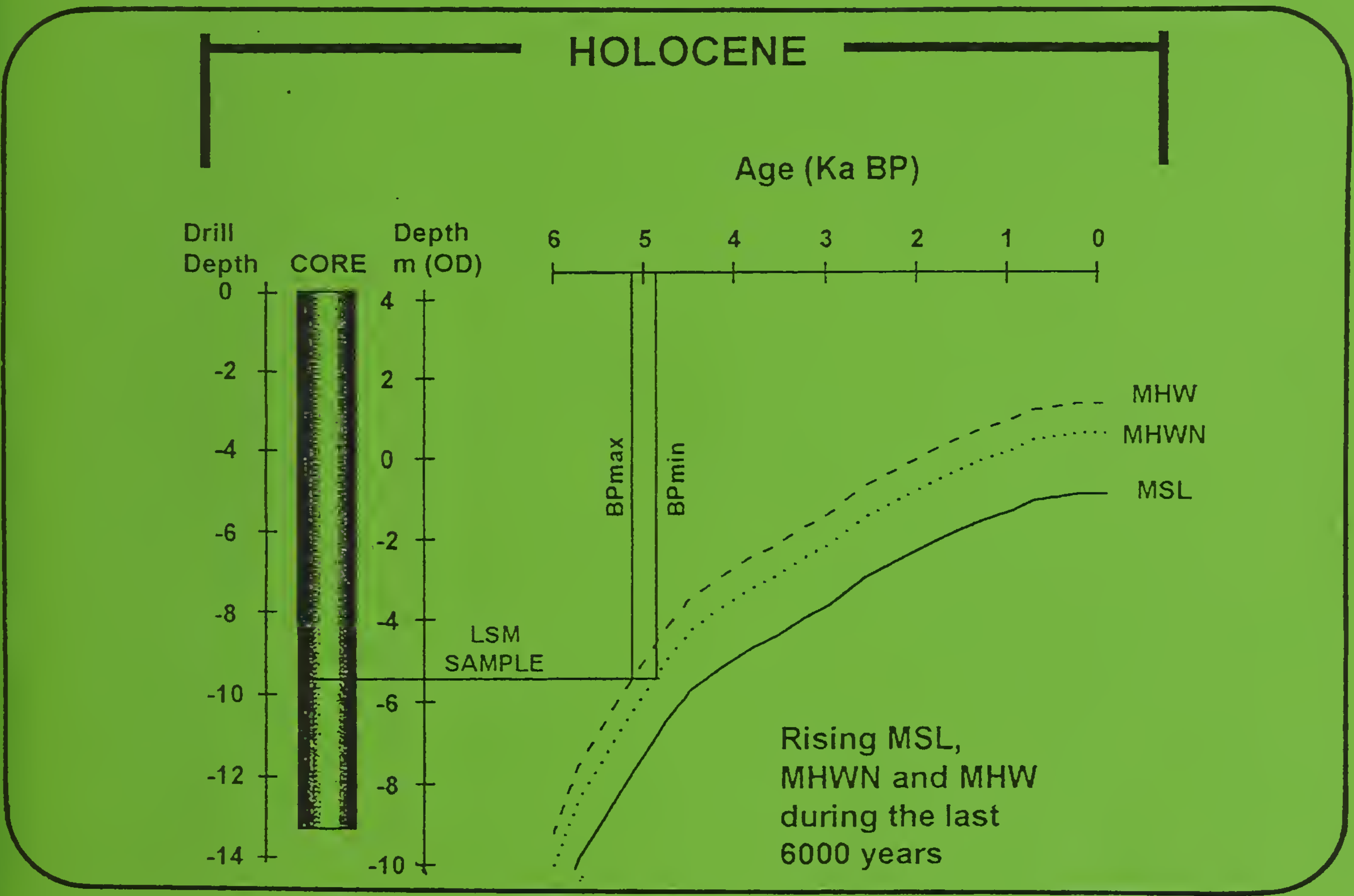
ISSN0143-9286

BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

(FOR ARTICLES ON THE GEOLOGY OF EAST ANGLIA)

NO.46

for 1996



PUBLISHED 1998

CONTENTS INCLUDE

Pleistocene mass movement

Periglacial ice-wedge casts

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BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

No. 46 (for 1996) Published 1998

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EDITORIAL

Bulletin No. 46 concentrates on the Pleistocene to Recent geology of the region. Three of the papers (West, Boomer and Funnell and Boomer) broadly fall into the category of environmental geology. This term is currently very popular and in part reflects the need for geologists to show that their science has direct relevance to humans and their activities. Environmental geology is particularly pertinent to the East Anglian region where sea-level rise and coastal erosion are very real issues impacting the populace. Of course we must not overlook more traditional aspects of our science, and the paper by Fish *et al.* uses observational data to interpret wedge casts in the Trimingham cliffs as permafrost features and not water escape structures.

I am short of material for issues beyond Bulletin 47 and welcome the submission of papers on any aspect of East Anglian geology. I hope to publish Bulletin 47 in the Spring of 1998 which will bring the publication schedule up to date. Finally, I would like to thank the authors who contributed to Bulletin 46 for their patience while I assembled the final copy.

INSTRUCTIONS TO AUTHORS

If possible, contributors should submit manuscripts as word-processor print out accompanied by a disk copy. We can handle most word-processing formats although PC Word, WordPerfect or ASCII files are preferred. In addition we accept typewritten copy and will consider legible handwritten material.

It is important that the style of the paper, in terms of overall format, capitalisation, punctuation, etc. conforms as strictly as possible to that used in Vol. 41 of the Bulletin. Titles and first order headings should be capitalised, centred and in bold print. Second order headings should be centred, bold and lower case. Text should be 1½ line spaced. All measurements should be given in metric units.

References should be arranged alphabetically in the following style.

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STEERS, J.A. 1960. Physiography and evolution: the physiography and evolution of Scolt Head Island. In: Steers, J.D. (ed.) *Scolt Head Island* (2nd ed.), 12-66, Heffer, Cambridge.

BLACK, R.M. 1988. *The Elements of Palaeontology*. 2nd Ed., Cambridge University Press, Cambridge. 404pp.

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The editors welcome original research papers, notes or comments, and review articles relevant to the geology of **East Anglia** as a whole, and do not restrict consideration to articles covering Norfolk alone. All papers are independently refereed by at least one reviewer.

A POSSIBLE LATE PLEISTOCENE MASS MOVEMENT IN FENLAND WITH AN ASSOCIATED MEDIEVAL SETTLEMENT: BURYSTEAD FARM, SUTTON, CAMBRIDGESHIRE

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ABSTRACT

A bulge on the slope forming the western edge of the Isle of Ely is interpreted as a result of a Late Devensian mass movement of Jurassic clays. It is associated with a long history of (human) medieval settlement, with a farmstead, chapel and moated site.

INTRODUCTION

The contours in the area of Burystead Farm, Sutton (G.R. TL 433789; Fig. 1), show a marked bulge of the 5m O.D. contour into the lower lying peatland of the Fens at the western edge of the Isle of Ely. Burystead Farm lies on a slight rise upon the bulge, while a moated site lies to the north-east between the farmstead and the rise to the plateau to the north-east. The two are separated from each other by the minor road leading from Sutton to Sutton Gault, the Bedford Rivers and the 100 Foot Washes. Higher contours show no such bulge, with the land north-eastwards rising gently to the plateau of the Isle of Ely at 20-25m O.D.

GEOLOGY OF THE STUDY AREA

The geological and soil maps of the area (B.G.S 1:50000 Series, Sheet 173 (Ely) - Gallois, 1988; Seale, 1974; 1975) show that Jurassic clays underly the bulge and the higher land to the north-east. To the south-west Ampthill Clay underlies Flandrian peat, and is overlain by Kimmeridge Clay as the slope rises to the east, with chalky boulder clay capping the sequence towards the top of the slope and on the plateau. The junction between the two Jurassic clays is recorded at a 1955 borrow pit 300m to the north of the bulge (Fig.1, D), and is illustrated in figure 13 of the British Geological Survey's area memoir (Gallois, 1988).

Bull. geol. Soc. Norfolk (for 1996) 46, 3-10 (Published 1998)

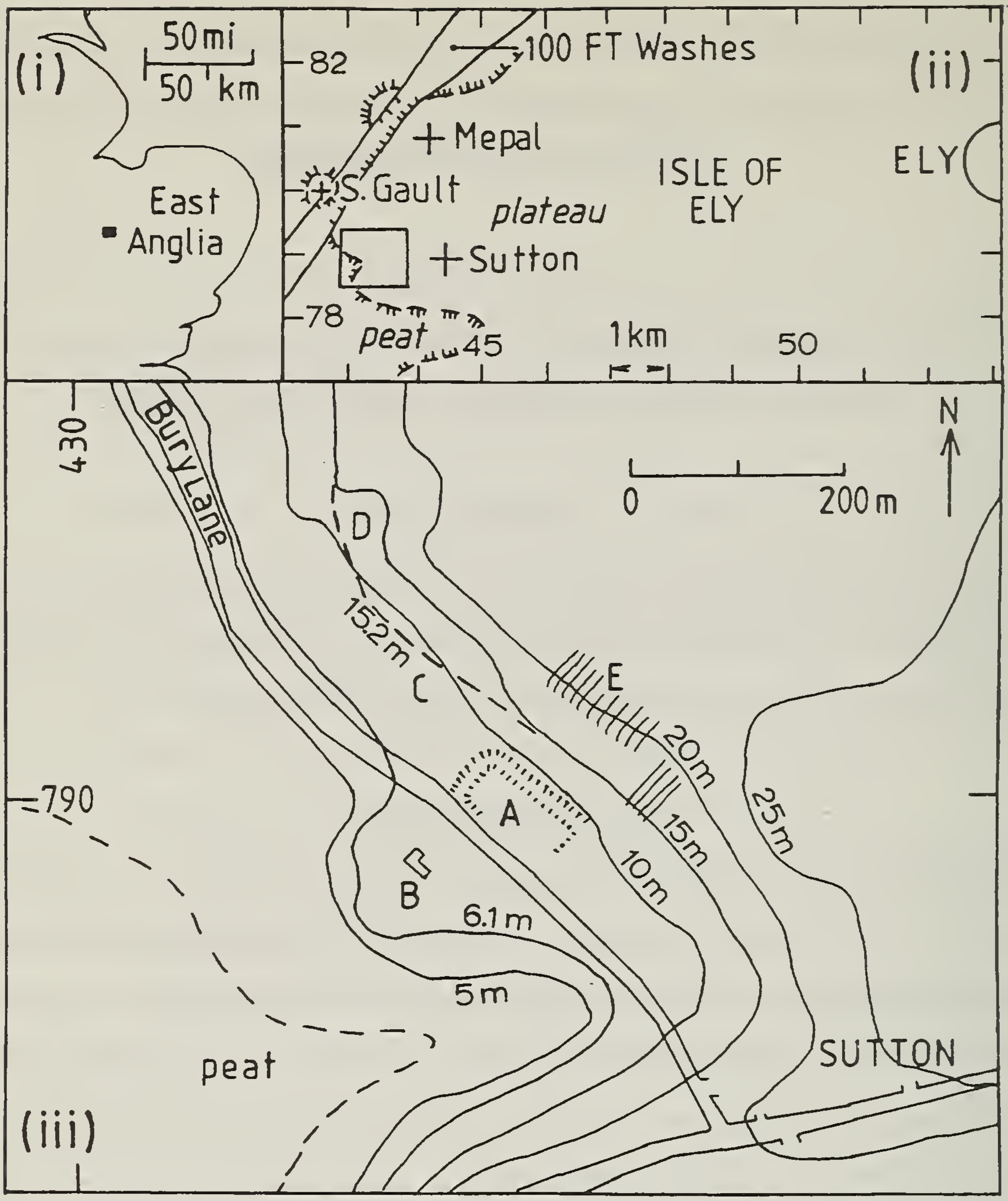

Fig. 1. (i) and (ii) location of study area at west margin of the Isle of Ely; (iii) the area surrounding Burystead Farm; A, moated site; B, Burystead Farm; C and D, borrow pits in Jurassic clay, the latter from 1955; E, ridge-and-furrow. The 6.1m and 15.2m contours are derived from the 1927 edition of the O.S. 6" map of the area.

Between Sutton and Mepal, seepages and a spring line are associated with mudstone bands in the Ampthill and Kimmeridge Clays (Gallois, 1988). Seale (1975) in the Soil Survey Memoir comments that "Whitish tabular limestone (in the Kimmeridge Clay) .... forms a relatively resistant layer from which water seeps near the top of the steep slopes between Sutton and Sutton Gault." He also describes the immediate area in the following way: "In

places between Sutton and Sutton Gault the Sutton ridge plunges at about 10° into the fens. Wicken soils occupy the convex steeper parts of this slope, the junction with the Hanslope series upslope at approximately 15 or 18m O.D. being marked by the outcrop of a band of limestone about a metre thick. Much of this hillside is in permanent grass and rushes grow where water seeps out of the limestone." Astbury (1958) also makes reference to the frequency of springs on the Isle of Ely, presumably related to the mudstone/limestone bands.

Figure 2 is a plan of the area showing the position of levelled surface transects stretching from the plateau south-westwards to the bulge and peat fen. The form of each transect is shown in Fig. 3.

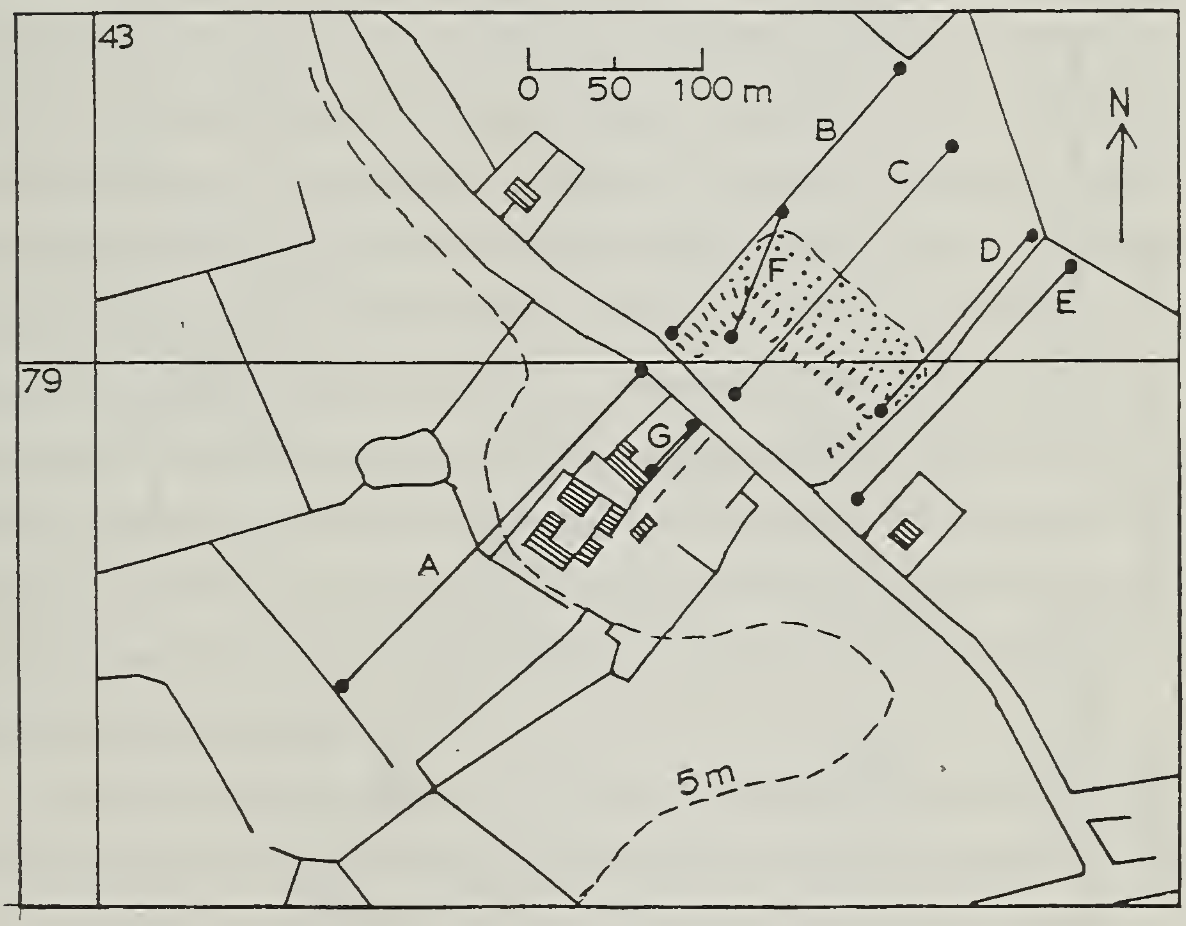

Fig. 2. Plan of levelled transects (A-G) in the area of Burystead Farm. The area of the back wall is dotted.

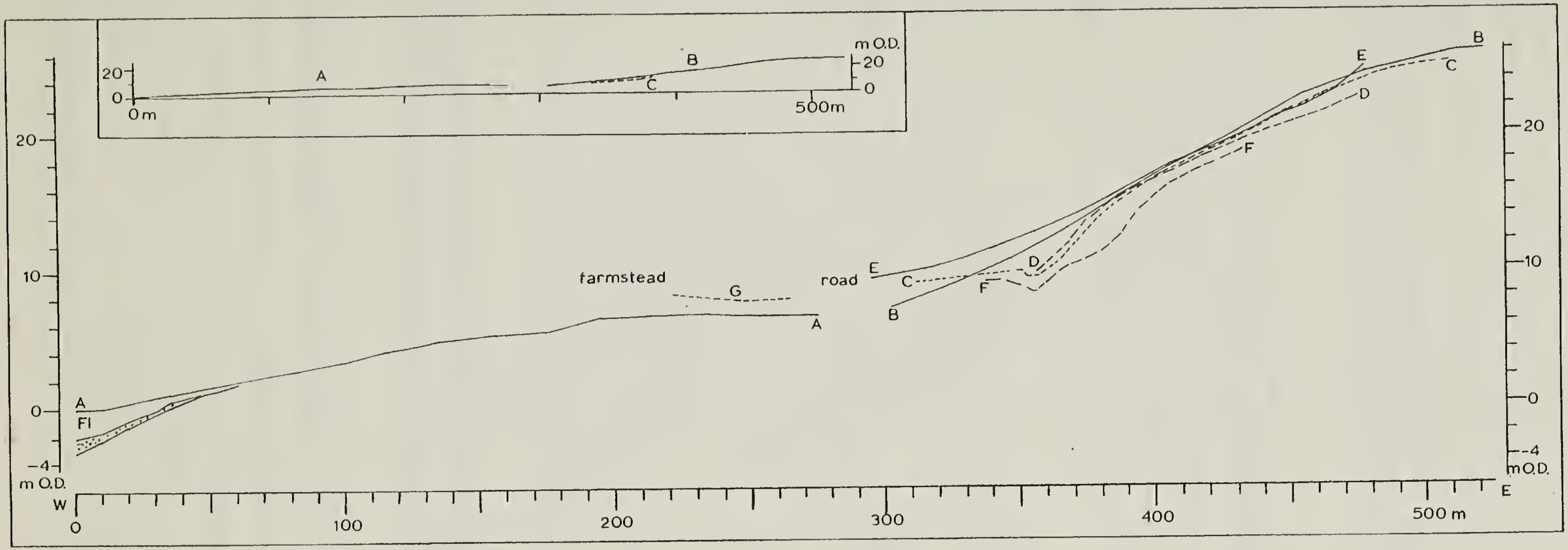

Fig. 3. Levelled transects (south-west to north-east) in the area of the mass movement. Vertical exaggeration X5; inset without vertical exaggeration. Fl = Flandrian sediments; stippled ornament = sand, silt and clay with flints.

INTERPRETATION

The interpretation of the bulge is confused by features of the human settlement, including the farmstead and moated site, by excavated ponds and/or depressions, by the more recent borrow pits to the north (Fig. 1, C & D), and by other possible physiographic changes since the time of its formation. However, the transects show an association of the bulge with a steep back wall to the east. Transect E, to the south of the moated site, shows a gentle continuous slope, as does transect B to the north, with a slope of c.10°. In contrast, transects C, D and F show the over-steepened slope (c.13°) of the back wall, directly north-east of the bulge in the 5m contour. The excavation of the moat may mask the presence of a cavity niche at the base of the backwall.

The form of the bulge suggests that it may result from flowage following a slope failure in the Jurassic clays. The transects suggest a bimodal flow, of a type illustrated and described by McRoberts & Morgenstern (1974), McRoberts (1978) and Harris (1981). Such flows show a relatively steep headscarp and a low-angle tongue. According to Harris (1981), they occur only in the presence of permafrost. The initiation of flows or slumps may begin with the exposure of permafrost after a slope failure, the melting of which results in the development of high pore-water pressures in fine sediment on the slope. Subsequent freeze/thaw processes, leading to enhanced pore pressures, leads to movement of fines even on low angle slopes, giving tongues spreading to lower ground from upslope failures. However, where seepages and springs are present, as in the Burystead area at present, the failure may have been encouraged or initiated by local hydrological factors, aided by freeze/thaw processes as the tongue developed. The presence of permafrost may not be certainly indicated, but at least seasonal freezing is suggested.

It seems probable that the age of the mass movement is Late Devensian - i.e. in the later part of the last cold stage of the Pleistocene - for the following reasons. The widespread Block Fen Terrace at c. 0m O.D. lies 3km to the west, and is the product of a gravel aggradation of the River Great Ouse at a time when this river followed a course northwards to The Wash between the Chatteris 'island' and the Isle of Ely (West, 1987; West, *et al.*, 1995). Radiocarbon dating of plant remains from sediments below the gravel of the Terrace gave a result of 18,550 (+1810 / -1470) radiocarbon yrs BP (Q-2657, West *et al.*, 1995), so the Block Fen Terrace is younger than this. Subsequently, and before about 11,000 B.P., the Great Ouse

abandoned its northwards course and turned to the west of Chatteris, cutting a deep channel which dissected the Block Fen Terrace gravel. This channel was later filled with organic sediment as the sea level rose in the Flandrian (Holocene), with peat and alluvium eventually spreading across the Fenland basin (Waller *et al.*, 1994). Thus at Haddenham, 1.4km to the south-west, a borehole in the channel of the Great Ouse showed some 4.5m of Flandrian organic sediments overlying Late Devensian late-glacial sediments at the base of the channel, junction at -5m O.D. (Waller *et al.*, 1994).

The typical Fenland Flandrian succession of a lower peat, fen clay and an upper peat was recorded in boreholes overlying the slope of the bulge at the south-west end of transect A, south-west of Burystead Farm (Fig. 3), demonstrating the pre-Flandrian age of the mass movement. These boreholes showed that sand, silt and clay with flints intervened between the Flandrian sediments and stiffer stoney clay, indicating reworking of the Jurassic sediment and fluvial activity before Flandrian organic sediments were formed.

It is possible that the mass movement at Burystead Farm is indirectly associated with the down-cutting of the Great Ouse channel, in which case it would have occurred at a time later than the formation of the Block Fen Terrace, but earlier than the Devensian late-glacial and the Flandrian. It is also possible, but perhaps less likely, that the slope failure is associated in part or whole with an earlier time of about 18,500 radiocarbon years BP, before the deposition of the Block Fen Gravel, when a lake formed in southern Fenland as a result of blocking of The Wash by the Late Devensian ice sheet. The formation of this lake, at a time of cold climate, was accompanied by massive wasting of clay slopes through thermokarst processes, especially thermal erosion, which resulted in the formation of the great embayments cut in the margins of the Isle of Ely and elsewhere on the Fenland margin (West 1991; 1993). At Somersham, 6 km to the west, a mass movement of Jurassic clay into the lake at this time is recorded (West, 1993). In this case, the cutting of the Great Ouse channel would be later than the mass movement. However, whatever the timing, there is evidence for permafrost at times during the whole period, before, during and after the deposition of the Block Fen Gravels, since large thermal contraction cracks are present below, in, and penetrating, the gravels. So the conditions promoting mass movement were present over a long period.

Whatever the timing, the mass movement at Burystead appears to be a bimodal flow or ground-ice slump, encouraged by the seepages and local spring line described above. A further point is that the presence of herbaceous vegetation, known to exist at the time from the fossil

record, would not have been any considerable hindrance to slope failure under the prevailing hydrological conditions.

HUMAN SETTLEMENT

The surface of the bulge shows a complex human settlement history. This includes the Burystead Moat site (Fig. 1A) and the Burystead Farm (Fig. 1B) (Salzman 1948; 1953). The latter incorporates a late 13th century - early 14th century chapel thought to have been associated with a former monastic grange (Historic Buildings Record, Cambs. C.C (1952); Pevsner, 1954). The mass movement must also be older than the east-west aligned ridge and furrow of the plateau (Fig. 1, E), which ends near the lip of the steep headscarp, with the pattern showing the characteristic bend towards the south. There are also further artificial features around the precincts of the chapel, including a possible precinct boundary to the north and west, and possible fishponds to the south which are enclosed between the bulge and the ridge on which Sutton stands.

Human settlement on the bulge surface took advantage of the slight rise above the general peatland level, with the chapel and farmstead on the highest ground, and the moated site near the headscarp of the mass movement. In some ways, the chapel site resembles the Priory site at Catley (G.R. TF119557), on a slight rise near the Fen margin in the River Witham valley to the north, seen in aerial view to show a possible precinct moat and fishponds (Knowles & St. Joseph, 1952). The moated area, between the chapel and farmstead area and the headscarp of the mass movement, adds additional interest to the site. Excavation may throw light on the history of the settlement and its associated features, which present an interesting combination of geology and archaeology.

ACKNOWLEDGEMENTS

I am grateful to K. Robinson for access to the area, to A. Taylor and C. C. Taylor for their advice on archaeological matters, to J. N. Hutchinson for advice on mass movements, and to S. Boreham and T. Chapman for assistance in the field.

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A PERIGLACIAL COMPOSITE-WEDGE CAST FROM THE TRIMINGHAM AREA, NORTH NORFOLK, ENGLAND

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ABSTRACT

A periglacial composite-wedge cast is described from Trimingham, north Norfolk, in deposits of the Cromer Forest Bed Formation, beneath Anglian glaciogenic sediments. The form and composition is taken to indicate that the feature formed in a periglacial climate, by a process of thermal contraction, deposition of wind-blown sand and snow, and subsequent ground-ice melt. The observation indicates that not all wedges observed in these sediments are water-escape structures as proposed by Worsley (1996) and confirms the existence of permafrost during the deposition of the Cromer Forest Bed Formation.

In a recent article, Worsley (1996) advises caution in interpreting wedge forms as ice-wedge casts, noting that many such features reported from the British Isles have been formed by soft-sediment deformation during water escape, rather than under periglacial conditions. He gives an example of this problem with a wedge from pre-glacial sediments at West Runton, north Norfolk. While fully accepting the principles and caution set out by Worsley, this report describes and discusses a wedge exposed during November 1996 at Trimingham, north Norfolk (TG 279 391) which displays properties typical of periglacial processes and confirms that many of the features recorded in north Norfolk, do indeed provide evidence of pre-glacial permafrost within the region (West, 1980).

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The feature under consideration is developed in medium fine white sands which comprise the upper part of the Cromer Forest Bed Formation (Fig. 1) (West, 1980, p 56, fig. 29). These sands are attributed by West (1980, p 56) to the early Anglian fluviatile sedimentation and the wedge is therefore in the position of one of the short-lived fragments of the Barham Soil developed in aggrading alluvial surfaces in north-east Norfolk (Rose et al., 1985, figs. 9.2, 9.3, 9.4). The sands are located beneath Anglian glaciogenic sediments.

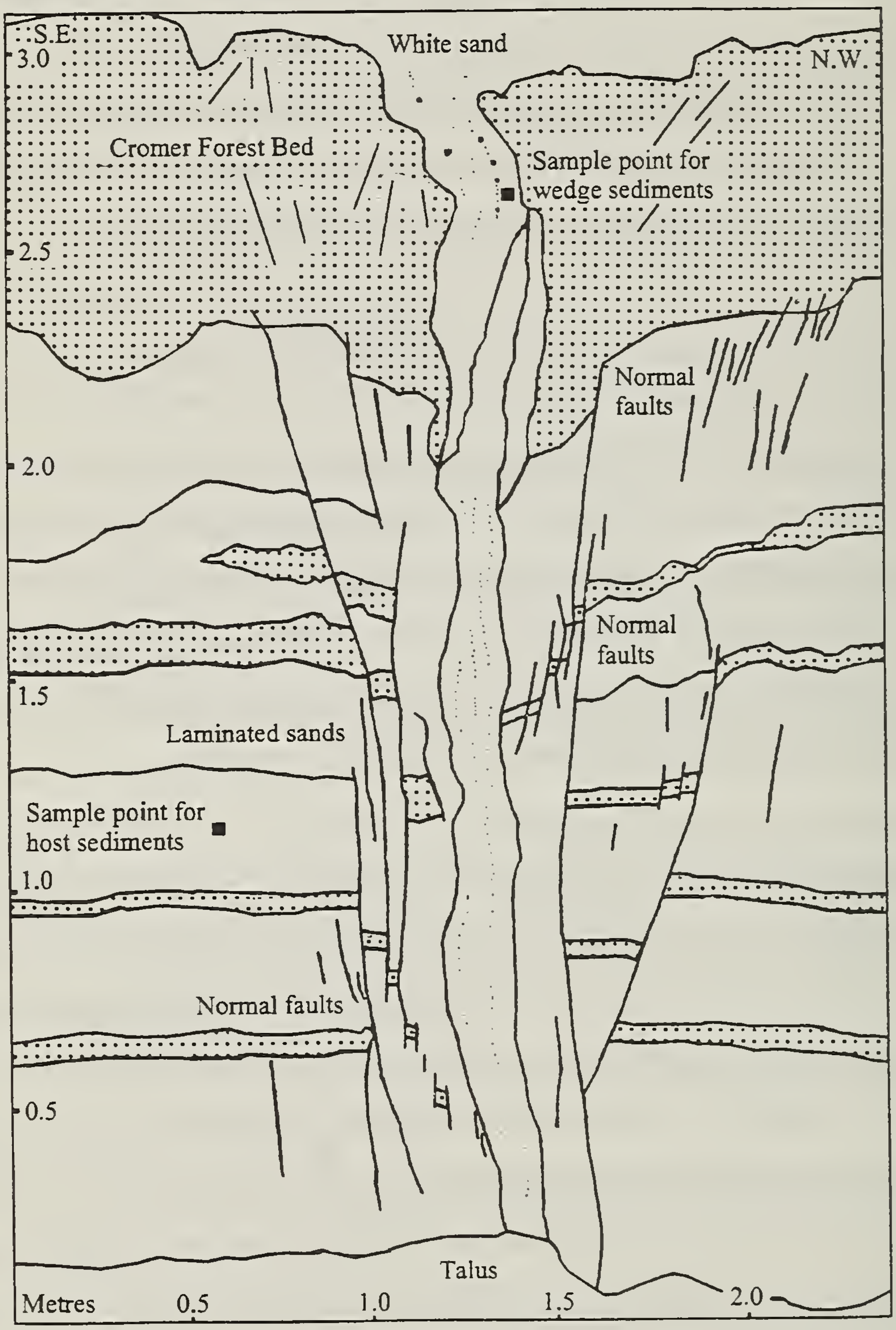

Fig. 1. The wedge feature at Trimingham, north Norfolk..

The wedge at Trimingham is almost 2 m long and up to 15 cm wide (0.75 m wide including external faults (Fig. 1). It consists of a vertical body of well sorted sand bounded by normal faults aligned parallel to the edge of the infill, and extending back into the cliff as part of a linear feature. No other wedges were visible on the occasion of the discovery, but adjacent sections of the cliff were obscured by sediment falls and similar wedges have been recorded in the area by one of the authors (JR) on previous occasions and by West (1980, fig. 29). Particle size analysis shows that the infill sediment is very similar in size range to the adjacent host material, but is slightly better sorted (Fig. 2) and contains wind-polished clasts.

The feature is interpreted as a composite-wedge (Black and Berg, 1966; Kolstrop, 1986, 1987; Gożdzik, 1973). The fill of well-sorted, vertically laminated sands with wind polished pebbles is typical of a sand-wedge, and the downward faulting of the host sediment is commonly associated with ice-wedge casts, developing in response to the void created by the melt of ground-ice or snow (Black and Berg, 1966; Harry and Gozdzik, 1988; Ballantyne and Harris, 1994; Murton, 1996). Clearly, a void has existed to receive aeolian sediment, and the host

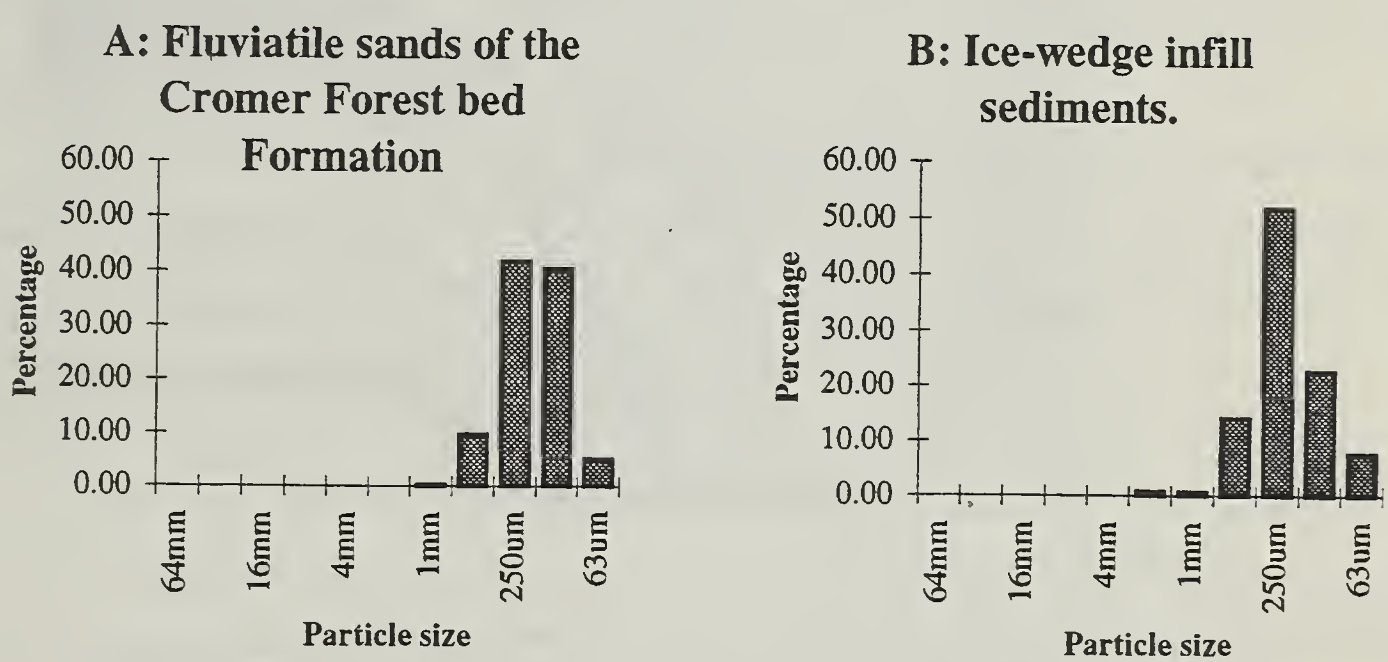

Fig. 2 Particle size distribution of the wedge infill and host sediments at Trimingham, north Norfolk.

Fig. 3. An ice wedge cast from the Leipzig area of Central Germany (from Eissmann and Wimmer, 1996).

material has experienced a change of phase enabling normal, gravitationally-driven faulting to take place. There is no evidence of lateral stresses associated with the growth of ground-ice (Black, 1976). On the basis of these observations it is considered that the feature at Trimingham formed in a periglacial environment as a ground contraction crack filled with sand blown in from the surrounding area, together with some snow, to form a composite-wedge. Subsequent thaw of the ground-ice, and snow within the wedge, perhaps due to climatic warming or fluvial activity in the periglacial environment, caused faulting at the edges of the wedge as the host sediments settled. The sand fill was, however, little modified and the vertical laminations within it were preserved. White, bedded sands overlying the wedge collapsed into the upper part of the void left by the ice suggesting that fluvial sedimentation had occurred between the formation of the composite-wedge and the melt of the ground-ice (Vandenberghe, 1983). In all respects the feature differs from the West Runton example cited by Worsley (1996) which is simply a wedge-shaped disturbance, lacking faulting and allochthonous infill.

The feature described at Trimingham displays many similarities to a wedge from Central Germany (Fig. 3), described and interpreted by Eissmann and Wimmer (1996) as being of periglacial origin, on the basis of up turning of host sediment, gravel infill and linear continuity through the exposure. The wedge is also very similar to that recorded by West (1980, plate 7) at West Runton, east of Woman Hythe Gap, also in early Anglian fluviatile sands (bed d) underlying Anglian till.

Worsley (1996) makes a valid point in cautioning field workers not to automatically interpret wedge features as being periglacial in origin. In this case, however, the evidence presented supports the observation of complex pre-Anglian periglacial phenomena in north Norfolk, which can be differentiated from soft sediment deformations.

ACKNOWLEDGEMENTS

P.R.F. acknowledges useful discussion with Drs Julian Murton and Colin Whiteman, and all authors wish to thank Dr. Charlie Harris for useful comments when reviewing the paper. With respect to R.J.O.H. this paper is published with permission of the Director, British Geological Survey (NERC).

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THE RELATIONSHIP BETWEEN MEIOFAUNA (OSTRACODA, FORAMINIFERA) AND TIDAL LEVELS IN MODERN INTERTIDAL ENVIRONMENTS OF NORTH NORFOLK: A TOOL FOR PALAEOENVIRONMENTAL RECONSTRUCTION.

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ABSTRACT

The relationship between the elevation of inter-tidal environments and floral and faunal distribution has been studied in a number of organisms. Two groups in particular (Ostracoda and Foraminifera) are particularly notable since they produce carapaces or tests which generally fossilise, thus providing an environmental record within the sediment. These two groups belong to the meiofauna (generally between 100 and 2000µm long) and as such are often recorded in high numbers from relatively small volumes of sediment. This makes them particularly useful for statistical analysis. As with other groups, species distributions are dependent upon elevation which reflects tidal inundation over the lunar cycle. By studying the relationship between ostracod and foraminiferal distribution on modern saltmarshes and intertidal flats it is possible to reconstruct past changes in sea-level by detailing the fossil assemblages in sedimentary cores. The following pages outline the faunal scheme used as part of a multidisciplinary study on the Holocene evolution of the north Norfolk coast.

INTRODUCTION

This research arises from work undertaken during the NERC (Natural Environment Research Council) community LOIS (Land-Ocean Interaction Study) programme. Under the LOEPS (Land-Ocean Evolution Perspective Study) component of that programme a number of projects have investigated the Holocene evolution of key coastal areas of Eastern England. North Norfolk was identified as an important area from the work of Pearson (1986) and Funnell & Pearson (1989). In 1994 a multidisciplinary investigation into the Holocene history of the north Norfolk coastline was initiated with particular

Bull. geol. Soc. Norfolk (for 1996) **46**, 17-29 (Published 1998)

emphasis on the sediments which were deposited along the coast during the last 8000 years of rising sea level. The emerging results of that project have been underpinned by detailed biofacies analyses of the Holocene sediments and the present paper outlines the background to these biofacies studies.

The Holocene sediments of north Norfolk are almost exclusively intertidal and can be separated into coarse-grained, high-energy barrier facies (sands and gravels) and fine-grained, low-energy back-barrier facies (saltmarsh/channel muds and silts). Many of the fine-grained sequences are quite homogeneous with little sedimentary evidence of environmental changes. Since we know that these sediments could have been deposited in a number of inter-tidal environments it was important to study the microfossil assemblages of these intervals in detail and thereby make confident palaeoenvironmental reconstructions.

BIOFACIES RELATIONSHIP TO TIDAL LEVELS

Previous studies have shown that the vertical zonation of living intertidal plant (e.g. Rodwell, 1995) and invertebrate species (e.g. Horne & Boomer, 1997) is related to local tidal conditions, although the absolute elevation range for a given species will vary from area to area (Scott & Medioli, 1978). The parameters which control distribution are not yet fully understood although the frequency of tidal inundation, tidal range, geomorphological setting, sediment supply and predation/grazing pressures are all certainly important. In an attempt to reconstruct Holocene intertidal environments it is necessary to establish the present day relationship between biological populations and tidal levels in the study area. This makes a primary assumption that tidal, climatic and geomorphological conditions have remained relatively constant during the past 10 000 years. Although these parameters will have varied during the Holocene the results of present study should still be applicable to past depositional environments.

Previous studies investigating the vertical zonation of intertidal invertebrates, levelled in reference samples to a known standard (ordnance datum in the case of the UK). In the context of the present work, the aim of the biofacies studies is to reconstruct changes in the environment of deposition during the Holocene. This study is based upon the conclusions of Pearson (1986) later published in Pearson *et al.* (1990) which related foraminiferal assemblages to the modern intertidal environments of north Norfolk; the present work confirms Pearson's observations. Neither Pearson's reference sets

(foraminiferal assemblages) nor those of the present study were levelled in to an absolute elevation. It must be stressed that the samples were taken to characterise particular *environments* (sediment type, macrofloral assemblages) rather than precise *elevations*. It is possible to translate environments into a tidal range, for example, upper saltmarsh accumulates between highest astronomical tide (HAT) and mean high water neaps (MHWN) along this coast and the precise elevation of these levels can be determined from the admiralty tide tables.

FORAMINIFERA AND OSTRACODA

Foraminifera and Ostracoda are easily recovered from Recent and Holocene sediments by washing the material through a series of sieves, usually between about 100 and 2000µm. The residue is then examined under a binocular microscope to reveal the microfossil assemblages.

The distribution of saltmarsh foraminifera (single-celled testate organisms) is known to be strongly correlated to tidal levels and, therefore, flooding frequency (Scott & Medioli, 1978, 1980). There is also evidence that salinity may have a marked influence in upper saltmarsh environments (De Rijk, 1995). The same saltmarsh associations are known world-wide (Murray, 1991; Phleger, 1970). The high abundance of foraminifera in intertidal sediments together with their strict vertical zonation makes them ideally suited for palaeoecological studies particularly in recording past changes in sea-level. Together with evidence from the ostracods these organisms make a powerful tool in the palaeoenvironmental reconstruction of saltmarsh environments.

A number of studies at the University of East Anglia have attempted to use foraminiferal assemblages to reconstruct Holocene intertidal and shallow sub-tidal palaeoenvironments in Norfolk (Coles, 1977; Godwin, 1993; Pearson, 1986). Boomer and Godwin (1993) integrated both foraminifera and ostracods into a reconstruction of two Holocene cores from eastern Broadland. Importantly, Pearson (*ibid.*) investigated both the modern surface sediments and Holocene sediments from the north Norfolk Coast establishing an important data set in the context of the current project. These studies, together with further research by the present author, have permitted a characterisation of modern intertidal environments through their invertebrate assemblages and so allow the reconstruction of Holocene palaeoenvironments. Table 1 presents a selection of samples

	UPPER MARSH				LOWER MARSH				UPPER MUDFLAT				LOWER MUDFLAT			
	IB75	BS2	BS4	BS5	IB54	IB62	IB69	IB76	IB68	IB48	IB50	IB52	IB79	IB80	IB81	IB82
Jadammina macrescens	42.2	43.1	56.5	48.9	3.8	2.5	1.1	4.2	8.5	1.0	0.4	0.2	1.6		0.6	
Trochammina inflata	57.1	56.4	42.2	46.0	0.6	1.5		5.6	3.8	0.5		0.7				1.0
Miliammina fusca			1.3	3.6	0.6											
***Haynesina* spp.**	0.7			1.4	53.7	37.1	47.3	23.7	73.7	87.3	92.3	94.6	0.0	3.8	45.2	1.0
Ammonia beccarii		0.6			6.2	11.1	17.8	2.4	2.6	2.6	1.5	0.9	70.5	88.5	19.9	80.2
Elphidium williamsoni					26.3	30.8	15.5	46.0	7.9	1.9	5.5	1.4	1.6		7.8	1.0
Quinqueloculina oblonga					8.3	16.4	6.8	1.0	0.3			0.7				
Elphidium incertum					0.6		0.4	0.3	0.3						0.6	
Elphidium gerthi						0.3									0.6	1.0
Miliolina subrotunda							0.8	0.3		0.7					0.6	
Elphidium earlandi							0.4	0.3	0.3							
Elphidium excavatum							6.4	2.8	1.2	3.6			16.4	3.8	21.7	8.3
Planorbulina mediterraneanensis							1.9	0.3	0.6	0.2		0.5				
Cibicides lobatulus						0.3	1.5		0.6		0.4		6.6	3.8	1.2	6.3
Elphidium magellanicum								1.0		1.9		0.9	1.6		1.2	0.0
Elphidium margaritaceum								11.8								
Quinqueloculina seminulum									0.3	0.2						
Massilina secans													1.6			1.0
Cyclogyra involvens															0.6	
Total number of specimens	294	181	232	139	339	396	264	287	342	418	273	427	61	26	166	96

Table 1. Selected samples from the four main intertidal environments identified within this study.

examined from the four main environments identified in this study. The foraminiferal assemblages contained within them confirm the observations of Pearson (1986) and Pearson *et al.* (1990).

Previous research

Pearson (1986) investigated the foraminiferal composition of over 250 sediment samples taken from the modern intertidal environments of north Norfolk. Using statistical cluster analysis he was able to divide the samples into four distinct assemblages, Upper Marsh, Lower Marsh, Marsh Creek/Intertidal Mud and Intertidal Sand/Gravel assemblages (Pearson, 1986; Pearson *et al.*, 1990). Pearson was also able to distinguish these units sedimentologically and, where present, on the basis of the macrophyte assemblages. There are three points which limit the potential of the Pearson classification: 1) only percentage abundance data now exist so some records may reflect very low occurrences which is a potential weakness in statistical analysis; 2) his study did not include ostracods which are sensitive indices of salinity variation and; 3) he recognised only 16 foraminiferal species whereas the present study encountered almost twice that number.

In an attempt to address these shortcomings, and to better refine the zonation, a number of new samples were taken from modern intertidal environments on the Norfolk coast, concentrating on the fine-grained (muddy and silty) sediments. The samples were not levelled into ordnance datum for the purposes of this study, rather, they were taken to characterise particular sedimentary and ecological environments which can then be related to tidal levels. The individual environments are primarily defined by their plant associations. Samples were taken from the upper saltmarsh, lower saltmarsh, upper saltmarsh pans, lower saltmarsh pans, major intertidal creeks, smaller saltmarsh creeks and intertidal mudflats. Samples were also taken from the intertidal sands and sandy channels for comparison. From the fine-grained sediment samples it was possible to recognise four foraminiferal and three ostracod assemblages.

Foraminifera

Foraminifera may be epifaunal, infaunal or adherent: many are capable of movement via their pseudopodia which are also used to collect food. In the intertidal environment, microscopic algae (including diatoms) form a large part of their food supply (Murray, 1979).

Group	Wall Structure	Intertidal Ecology	Main Saltmarsh Genera
Rotaliina	Calcareous, hyaline	Marine - Brackish. Lower Saltmarsh	*Ammonia, Elphidium, Haynesina*
Miliolina	Calcareous, porcellaneous	Marine. Lower Saltmarsh	*Quinqueloculina, Miliolinella*
Textulariina	Detrital siliceous, agglutinating	Brackish - Marine. Upper - Lower Saltmarsh	*Jadammina, Trochammina*

Table 2. The three major groups of foraminifera found on British saltmarshes with details of their wall type, general ecology within the intertidal zone and the dominant intertidal genera.

Saltmarsh foraminifera can be divided into three broad groups based on the structure of their test, a feature that forms the basis of their taxonomic classification. The calcite walled forms comprise two sub-groups, 1) the Rotaliina possess radially arranged calcite crystals conferring a glassy or hyaline appearance to the test while, 2) the Miliolina have a lamellar arrangement of calcite crystals resulting in a porcellanous appearance. The third major group are the agglutinating walled Textulariina which construct their tests from locally available, generally siliceous, detrital material such as mud and silt grains agglutinated onto an organic lining. These three groups are summarised in Table 2 together with the main representative saltmarsh genera and their general ecology. It is clear from this table that the agglutinating walled foraminifera dominate the upper saltmarsh while the Rotaliina dominate intertidal mudflats. The lower saltmarsh environment generally comprises an intermediate assemblage.

Four foraminiferal assemblages are recognised each comprising a number of species. This interpretation is complicated by the natural occurrence of individual species in more than one of the assemblages and also the *post-mortem* transport of shells and tests. The autochthoneity of a sample is assessed from characteristics such as size-sorting, population structure and preservation. In all cases the interpretations are based on what is considered to be the *in situ* component. The four foraminiferal assemblages recognised,

their inferred environment and constituent taxa are summarised below. The interpretation of each assemblage has been determined through a number of criteria, largely based on the work of Pearson (1986), but validated by the recent sampling programme. The four assemblages are thought to equate to the following tidal levels.

USM (Upper Saltmarsh)	HAT-MHW
LSM (Lower Saltmarsh)	MHW-MHWN
UMF (Upper Mudflat)	MHWN to about MSL
LMF (Lower Mudflat)	MSL? to about MLWN

Upper Saltmarsh assemblage (USM)

All of the species in this assemblage are agglutinating taxa. The assemblage is dominated by *Trochammina inflata* and *Jadammina macrescens* which together usually constitute at least 99% of the total number of individuals. A sub-assemblage, UUSM (uppermost upper saltmarsh), has also been recognised at the highest part of the upper saltmarsh where tidal inundation is least frequent. Samples from this sub-assemblage yielded few or no foraminifera. Towards the top of the saltmarsh assemblages tend to become dominated by *T. inflata.*

Lower Saltmarsh assemblage (LSM)

This assemblage is in some ways transitional between upper saltmarsh and upper mudflat. It is in this environment that the highest abundances of *Miliammina fusca* and *Elphidium williamsoni* are recorded, commonly reaching 20% each. The assemblages generally comprise a mixture of agglutinating and calcareous walled taxa.

Upper Mudflat assemblage (UMF)

Haynesina germanica, Haynesina depressula and *Ammonia beccarii* are generally the most dominant species in this assemblage, with the exact composition dependent on a number of factors, particularly salinity variation. These species together with some agglutinating forms and minor constituents such as *Quinqueloculina oblonga* generally constitute up to 95% of the assemblage.

Lower Mudflat assemblage (LMF)

This assemblage is similar to the upper mudflat except that the *Ammonia* and *Haynesina* species generally constitute a smaller percentage of the total, with species of *Elphidium* (except *E. williamsoni*) and the Miliolina becoming much more common. Shelf genera (e.g. *Asterigerinata, Cibicides, Glabratella, Lagena* and *Planorbulina*) are common in small numbers in this assemblage indicating the proximity to more open water conditions.

Ostracoda

The Ostracoda (ostracods) are a group of small Crustacea (generally 0.3-1.5mm long) which inhabit a wide range of marine and non-marine environments. Although usually numerically subordinate to other meiofauna such as nematodes, copepods and foraminifera they sometimes achieve very high population densities (Heip, 1976). Ostracod distribution is limited to the lower saltmarsh environments, drainage channels, tidal creeks and upper saltmarsh saltpans: they do not occur living on the upper saltmarsh surface (i.e. above MHW). They can survive temporary emergence in residual pools, among damp seaweed/algae or within the unconsolidated upper layers of wet sediments. In contrast, some foraminifera are capable of withstanding sustained emergence on the upper saltmarsh surface. Ostracod species distributions are primarily controlled by salinity variation, temperature, oxygen availability and substrate type. While the foraminifera may be used in the first instance to indicate the tidal level of deposition, the ostracods can provide both confirmation of the foraminiferal assemblages and also an assessment of salinity levels and variation. Three ostracods assemblages have been identified in this study and are summarised below together with their most important constituent taxa.

Brackish Ostracod assemblage

Species in this assemblage live in waters where salinity is generally maintained below 'normal' seawater (typically 10-25‰). These environments are usually saltpans, enclosed ditches or other water bodies which are only flooded on highest spring tides. There are few significant freshwater inputs to the north Norfolk Coast so estuarine conditions are not relevant in the context of this study. Absolute salinity levels can fluctuate widely and may sometimes exceed those of seawater; *Cyprideis torosa* is known to tolerate salinities of 1-140‰ (Athersuch *et al.*, 1989). The main species in this assemblage are *Cyprideis torosa, Leptocythere castanea, Leptocythere porcellanea, Loxoconcha elliptica.*

Lower Saltmarsh/Euryhaline Ostracod assemblage

The species in this assemblage are capable of withstanding significant environmental changes in terms of both tidal inundation and salinity (salinity range experienced is approximately 15-35‰) on a daily basis. They are characteristic of British lower saltmarsh environments (Horne & Boomer, 1997) and the main channels which drain the upper saltmarsh. These environments are flooded on all but the lowest neap tides. Typical species include *Hemicythere rubida, Leptocythere lacertosa, Hirschmannia viridis, Leptocythere psammophila, Cytherois fischeri, Leptocythere baltica* and *Elofsonia baltica*.

Marine Ostracod assemblage

This assemblage contains species which are not tolerant of salinity reduction and most of them are common in the sublittoral zone of the British coast (Athersuch *et al.*, 1989). These taxa are often recorded as juveniles in intertidal sediments, and then in low abundance. In such cases they are most likely allochthonous, having been brought into the intertidal zone due to wave or storm surge action. Their presence is considered to indicate deposition around mean sea level (MSL). The common constituent taxa are *Pontocythere elongata, Semicytherura* spp., *Leptocythere pellucida, Loxoconcha rhomboidea* and *Hemicythere villosa*.

AN APPLIED EXAMPLE FROM NORTH NORFOLK

During the course of the north Norfolk Coast project a number of deep cores (up to 20 metres deep) and shallower auger sections (up to 8 metres) were taken through the Holocene deposits. Table 3 details the foraminifera and ostracods recovered from one augered section, Core 96/17. This core was taken at the seaward edge of the lower saltmarsh at Stiffkey and penetrated almost five metres of intertidal silts (National Grid Ref. TF 96474489). The percentage of each foraminiferal and ostracod assemblage is summarised for each sample and these are presented in Figure 1.

The foraminifera show quite clearly that an upper saltmarsh environment persisted at this locality for much of the period of deposition of these sediments. The ostracods are recovered in low abundances in this core but generally support observations based on the foraminifera. One sample at about 2.00 metres depth indicates a relative increase in lower

NNC96/17 Biofacies Data (Number of specimens)

		FORAMINIFERA								OSTRACODA					
Drill depth (m)	Corrected depth (m OD)	*Jadammina macrescens*	*Trochammina inflata*	*Elphidium williamsoni*	*Haynesina germanica*	*Haynesina depressula*	*Ammonia beccarii*	*Elphidium excavatum*	*Elphidium magellanicum*	*Leptocythere porcellanea*	*Leptocythere lacertosa*	*Leptocythere baltica*	*Hemicythere rubida*	*Leptocythere psammophila*	*Pontocythere elongata*
0.57	1.93	1	1	7		4	4	1	7			5		3	1
0.75	1.75	128	72	4	1	3			4						
1.25	1.25	1	3												
1.80	0.70	177	12												
2.50	0.00	36	2	8	54	20	2		8	5	6		2		
3.60	-1.10	42	1								1				
4.75	-2.25	95	90	1	5	3			1						
Assemblage		USM		LSM	UMF			LMF		Br.	LSM/Eury.			Marine	

Corrected depth (m OD)	% Upper Saltmarsh foraminifera	% Lower Saltmarsh foraminifera	% Upper Mudflat Foraminifera	% Lower Mudflat foraminifera	% Brackish Ostracoda	% LSM/Euryhaline Ostracoda	% Marine Ostracoda	Environment
1.93	8	28	32	32	0	56	44	LSM - Euryhaline
1.75	94	2	2	2	0	0	0	USM
1.25	100	0	0	0	0	0	0	USM
0.70	100	0	0	0	0	0	0	USM
0.00	29	6	58	6	38	62	0	LSM - Euryhaline
-1.10	100	0	0	0	0	100	0	USM
-2.25	95	1	4	1	0	0	0	USM

Table 3. Foraminiferal and ostracod distribution in Core 96/17 from the Lower Saltmarsh at Stiffkey. An environmental interpretation is given in the right hand column. Each sample is 1cm thick with the depth given to the top of each sample.

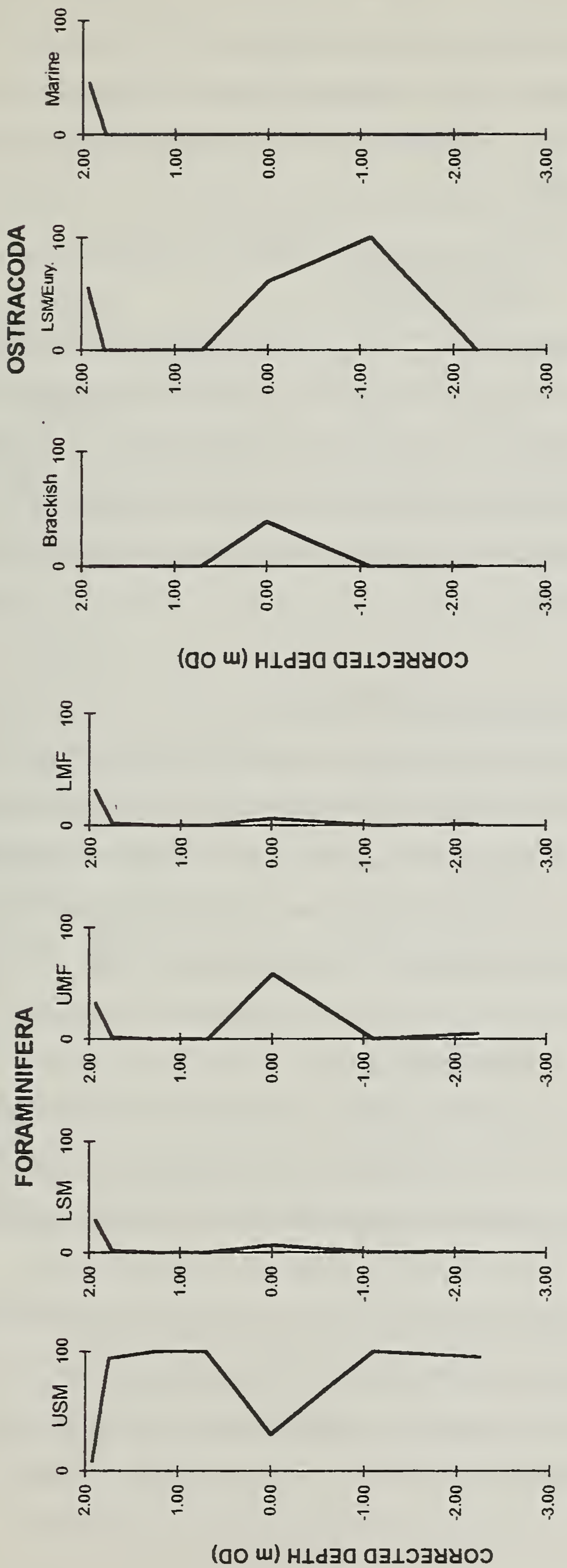

Fig. 1. Percentage abundance of the four foraminiferal and three ostracod assemblages in Core 96/17 from Stiffkey.

saltmarsh and upper mudflat species. This change must have resulted either from an increase in the rate of sea-level rise or a decrease in sediment supply. The uppermost sample indicates a further seaward shift in the environment of deposition with upper saltmarsh foraminifera reduced almost to zero. The assemblage comprises a mixture of lower saltmarsh and mudflat species indicating lower saltmarsh/euryhaline conditions much as we see at that locality today.

CONCLUSIONS

Four distinct foraminiferal assemblages (and one sub-assemblage) and three ostracod assemblages are recognised on the modern, fine-grained, intertidal environments of the north Norfolk Coast. These assemblages can be used to determine the environment (and therefore the tidal level) at which Holocene sediments were deposited based on their fossil microfaunas. The method has implications for studies of past changes in relative sea-level and sediment supply to this coastline.

ACKNOWLEDGEMENTS

The author wishes to thank the following co-workers on the NERC Special Topic No.32; GST/02/737, for their support in this research: Dr Julian Andrews, Prof. Brian Funnell and Dr Rina Jones (all UEA). Thanks also to Dr Greg Samways (UEA) for his invaluable assistance in the field. The author wishes to thank Dr Roland Gehrels (University of Plymouth) for his critical review of the paper. LOIS publication number 308.

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MICROBIOFACIES TIDAL-LEVEL AND AGE DEDUCTION IN HOLOCENE SALTMARSH DEPOSITS ON THE NORTH NORFOLK COAST

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ABSTRACT

Saltmarshes develop in close relationship to tidal-levels. Along the North Norfolk Coast, judged mainly on the basis of range relative to Ordnance Datum (OD), uppermost saltmarshes appear to range between Highest Astronomical Tide levels (HAT) and a mean between HAT and Mean High Water Spring (MHWS) levels (HAT-MHWS); upper saltmarshes between the HAT-MHWS mean and a mean between MHWS and Mean High Water (MHW) levels (MHWS-MHW); and lower saltmarshes between the MHWS-MHW mean and MHW levels. Since HAT, MHWS, and MHW tidal-levels are systematically related to OD, and also to Mean Sea Level (MSL), in a vertical sense, present and past MSL levels can be deduced from the occurrence of different types of saltmarsh (defined by microfauna).

Post-glacial changes in MSL affecting both the North Sea and the North Norfolk Coast have been modelled by Lambeck (1995), producing contour (isobase) maps relating to MSL along the North Norfolk Coast at 8000 and 7000 BP (years before 1950 AD). By comparison with the modelled history of sea-level rise in Fenland since 7000 BP the likely progress of MSL rise along the North Norfolk Coast since 7000 BP can be calculated.

Then, by: (a) deducing the tidal-level at which saltmarsh deposits in boreholes on the North Norfolk Coast were formed, and (b) using the Holocene time at which that tidal-level would have corresponded to the OD/MSL depth at which those deposits are now found, a provisional age relative to BP can be assigned.

Bull. geol. Soc. Norfolk (for 1996) **46**, 31-55 (Published 1998)

INTRODUCTION

In 1994 a programme of borehole drilling was commenced on the North Norfolk Coast to study Holocene coastal sediments, as part of the U.K. Natural Environment Research Council's LOIS (Land-Ocean Interaction Study) programme, designed to look at land-sea interactions, and in particular the LOEPS (Land-Ocean Evolution Perspective Study) component, looking at changes in coastal fluxes during the last 10,000 years. The potential of this coast for such a study had previously been revealed by Murphy and Funnell (1980, 1985), the theses of Allison (1985) and Pearson (1986), and the publications of Funnell and Pearson (1984, 1989) and Pearson, Funnell and McCave (1990).

Radiocarbon dating can be used to date such sediments. Radiocarbon dates for North Norfolk are contained in the theses of both Allison (1985) and Pearson (1986), and the publication derived from one of them (Funnell and Pearson, 1989). However, in spite of the potential accuracy and precision of the modern radiocarbon method, there are still problems associated with dating organic material by this method. For example: peat may contain detrital material from an older peat deposit; conversely roots may penetrate deep into the substrate introducing relatively younger carbon to these horizons. [It should be noted that radiocarbon ages BP (Before Present) are expressed in relation to 1950 AD as year zero, *not* to 0 AD.]

Tidal, elevation and age abbreviations used in this paper

HAT	=	highest astronomical tide
HAT-MHWS	=	*mean between* highest astron. tide - mean high water springs
MHWS	=	mean high water springs
MHWS-MHW	=	*mean between* mean high water spring - mean high water
MHW	=	mean high water
MHWN	=	mean high water neaps
MSL	=	mean sea level
LAT	=	lowest astronomical tide
CD	=	Chart datum (usually very close to LAT)
OD	=	Ordnance datum [Newlyn] (usually close to local MSL)
BP	=	Before present (1950 AD)

Additional methods of dating were undertaken as part of the LOEPS work on the North Norfolk coast. Matching the variation of the geomagnetic vectors recorded in the sediment cores to the known Holocene time-variation of the Earth's magnetic field has been undertaken at UEA (Dr. B.A. Maher and Dr. I. Boomer), luminescence dating of sand grains at Durham University (Mr. I.K. Bailiff and Dr. H. Roberts), and radionuclide mass accumulation 'dating' at Edinburgh University (Dr. G.B. Shimmield and Mr J. Smith). The samples for these dating methods came from selected cores, often specific to the methods chosen. Moreover, their application has been limited by both technical difficulties and cost.

Unfortunately, fewer than expected well-preserved and positioned peat deposits suitable for radiocarbon dating were found in the cores. Also considerable delays were experienced in processing those that were recovered. Therefore, their contribution to the age interpretation of the sediments in the earlier stages of our work was very limited. The need for an alternative dating method became apparent, and what was developed forms the basis of this paper. It is based on knowledge of:

(a) the fact that some inter-tidal sediment facies, particularly saltmarshes, can be identified by their microfossil content, which have a systematic relationship to tidal-levels that have specific heights relative to Ordnance Datum (OD) and mean sea level (MSL) values;

(b) MSL values over the last 7,000 years (Lambeck, 1995), relative to present-day OD (Figure 1 and Table 3). Although these values cannot possibly be regarded as absolutely accurate or precise, they are likely to have sufficient general consistency for the present purpose;

If we: (i) know the present-day OD depth from which our sample is obtained;

(ii) can deduce the past inter-tidal facies levels within which it was formed;

(iii) can attribute those past inter-tidal levels in terms of present-day OD level equivalents;

we can infer the age at which the deposit was formed.

There is no intention to claim that this is an absolutely accurate or precise method, or indeed more than a general and overall method. However we shall show our arguments and calculations in full, and show how this method of estimating ages in this sequence of

sediments appears to provide a method that actually works rather well in allocating a time framework to the sedimentary sequence.

SALTMARSH LEVELS RELATIVE TO SEA-LEVEL

By studying the foraminifera and ostracods preserved in inter-tidal sediments, it is possible to distinguish the sediment types and deduce the tidal-levels at which these sediments were deposited (Scott and Medioli, 1978, 1980). In particular it is possible to identify and distinguish past upper and lower saltmarsh deposits. North Norfolk Holocene coastal sections, interpreted using this method, have been published by Funnell and Pearson (1984, 1989). The vertical inter-tidal levels within which they accumulated can be inferred.

The species of foraminifera and ostracods that live in and on saltmarshes are systematically distributed relative to inter-tidal levels, and become preserved as microfossils in saltmarsh deposits. In doing so they preserve a record of the inter-tidal levels at which the saltmarshes accumulated. A considerable number of papers have been published on this general subject on a world-wide basis, but numerical tidal ranges and conditions vary in different parts of the world's oceans and marginal seas, so the results are only generally, rather than totally consistent. Fortunately, work along both the North Norfolk, and other parts of the East Anglian coast, provide useful reference information concerning inter-tidal levels for microfossils which are relevant in upper and lower saltmarshes regionally (Pearson, 1986; Funnell and Pearson, 1989; Brew, 1990; Boomer 1998 - this issue of BGSN). However, even these are not totally well enough studied at the present time for the present purpose. Until now it has been thought that on the North Norfolk Coast the upper saltmarsh extends from Highest Astronomical Tide level (HAT) to Mean High Water tide-level (MHW), and lower saltmarsh from Mean High Water tide-level (MHW) to Mean High Water Neap tide-level (MHWN), forming a clearly recognisable pattern.

However, the relationship between saltmarsh types and inter-tidal levels continues to be investigated. Brown and Gray (pers. comm.) have defined their Upper Saltmarsh as ranging between HAT and MHWS; their Middle Saltmarsh as ranging between MHWS and MHW; and their Lower (Pioneer) Saltmarsh as ranging between MHW and MHWN. The Upper [Salt]Marsh of Pearson *et al.* (1990) is therefore likely to be equivalent to the Upper and Middle Saltmarsh of Brown and Gray (pers. comm.), and Pearson *et al's*

Lower (Salt)Marsh to the Lower (Pioneer) Saltmarsh of Brown and Gray (pers. comm.). The Saltmarsh of French and Spencer (1993) appears to correspond to most, but not necessarily all, of the vertical range allocated to the Upper and Middle Saltmarsh by Brown and Gray.

Having defined microfossil biofacies corresponding to the upper and lower saltmarsh, it is necessary to relate those biofacies and their related saltmarshes directly to OD rather than tidal levels. It is the inter-tidal conditions, rather than the OD levels, that are most important in defining the saltmarsh and biofacies types. It is possible, however, to define the relationship between inter-tidal facies and OD levels on the North Norfolk Coast.

PRESENT-DAY TIDAL RANGES NEAR TO THE NORTH NORFOLK COAST

Tidal ranges along the North Norfolk Coast are actually difficult to determine in both a general and specific fashion. However, new results from the Wash have provided a novel way of looking at the problem. Using satellite imagery Mason *et al.* (1997) have constructed a digital height map of the inter-tidal zone of the Wash, showing a range from +2.56m to -2.50m, with a mean value of +0.03m. This can be compared with the actual tidal data from the Admiralty Tide Tables for 1997 (1996) from Skegness to Hunstanton (Table 1).

It is clear from these comparisons that the range of 5.06m shown on the digital height map (Mason *et al.*, 1997) is slightly less than the average MHWS-MLWS range of 6.22m shown by the Admiralty stations around the Wash

No comparable digital height map is available for the North Norfolk Coast. However, the tidal levels can be compiled in the same way as they have been for the Wash (Table 2). This is only done for the range down to MSL, first to CD and then to OD, but also, because of our interest in the highest tidal range in relation to saltmarsh development, an estimate for HAT is also included.

It should be noted from Table 2 that the height of:

HAT appears to vary from +3.5 to +3.2m between Burnham and Blakeney

HAT-MHWS appears to vary from +3.20 to +3.00m

(Only between HAT and HAT-MHWS does the saltmarsh seem to have *upper* upper saltmarsh characteristics.)

Table 1. Current tidal ranges around the Wash. All values in metres relative to Chart Datum

	MHWS	MHWN	MSL	MLWN	MLWS
Skegness	6.9	5.3	4.01	2.5	0.9
Boston	6.8	4.8	3.31	1.8	0.4
Wisbech Cut	7.0	5.1	---	2.2	---
King's Lynn	6.8	5.0	3.56	1.8	0.8
Hunstanton	7.4	5.6	3.88	2.5	0.9
Immingham	*7.3*	*5.8*	*4.2*	*2.6*	*0.9*

[These figures have been obtained from the Admiralty Tide Tables for 1997 (1996). The MHWS, MHWN, MLWN and MLWS values have had to be calculated, by reference to the Immingham CD values on p.306. Most MSL values are given directly on p.306, but the MSL values for Immingham are given in Table V on p.xxxix. The *Immingham* figures have been added for reference at the base of the above listing. All tidal heights in this listing are referred to Chart Datum (CD), which generally corresponds to Lowest Astronomical Tide (LAT), but for Immingham LAT is shown in Table V on p.xxxix as +0.1m CD.]

	MHWS	MHWN	MSL	MLWN	MLWS
Skegness	3.15	1.55	0.26	-1.25	-2.85
Boston	3.93	1.93	0.44	-1.07	-2.47
Wisbech Cut	3.8	1.9	-	-1.0	-
King's Lynn	3.77	1.97	0.53	-1.23	-2.23
Hunstanton	3.65	1.85	0.13	-1.25	-2.85

[The OD values for CD for these stations are given in Admiralty Tide Tables for 1997 (1996, Table III, p.xxxiii).]

Table 2. Current tidal ranges along the North Norfolk Coast. All values in metres relative to Chart Datum

	HAT	MHWS	MHW	MHWN	MSL
Hunstanton	8.0	7.4	6.5	5.6	3.88
Burnham (O/S)	2.9	2.3	1.6	0.9	---
Wells (Bar)	6.60	6.0	5.4	4.8	---
Wells	4.1	3.5	2.75	2.0	1.19
Blakeney (Bar)	6.3	5.7	5.1	4.5	---
Blakeney	4.0	3.4	2.7	2.0	---
Cromer	5.8	5.2	4.65	4.1	2.80

[These figures, for the upper part of the tidal range only, have been obtained and calculated from the Admiralty Tide Tables for 1997 (1996) as for the Wash, but only for MSL and above. MHW has been calculated as the mean between MHWS and MHWN. HAT has been simply calculated by adding 0.6m. The difference between MHWS and HAT given in Table 5, p.xxxix for Immingham is 0.7m and for Lowestoft 0.5m, so an average of 0.6m has been applied to the North Norfolk Coast.]

	HAT		MHWS		MHW	HWN	MSL
Hunstanton	4.25		3.65		2.75	1.85	0.13
Burnham (O/S)	**3.50**	**3.20**	**2.90**	**2.55**	**2.20**	**1.50**	---
Wells (Bar)	5.85		5.25		4.65	4.05	---
Wells	**3.35**	**3.05**	**2.75**	**2.37**	**2.00**	**1.25**	0.44
Blakeney (Bar)	5.50		4.90		4.30	3.70	---
Blakeney	**3.20**	**3.00**	**2.60**	**2.25**	**1.90**	**1.20**	---
Cromer	3.05		2.45		1.90	1.35	0.05

[It is noticeable that the figures relating to Wells and Blakeney Bars are not consistent with the otherwise regular sequence of Burnham Overy Staithe, Wells and Blakeney values between Hunstanton and Cromer. Also the Wells MSL value relative to OD is appreciably different to the Hunstanton and Cromer values relative to OD. For the present purpose the first of these types of anomaly can be overlooked. For the second a simplifying assumption is made that in the past MSL was the same as its past OD-equivalent, because what we wish to do is estimate a past MSL (not specifically a past OD value) by deducing it from biofacies-defined past HAT/MHWS/MHW depths. Only the figures in bold are used in our calculations.]

The height of:

HAT-MHWS appears to vary from +3.20 to +3.00m between Burnham and Blakeney

MHWS-MHW appears to vary from +2.55 to 2.25m

(Only between HAT-MHWS and MHWS-MHW does the saltmarsh seem to have upper saltmarsh characteristics.)

The height of:

MHWS-MHW appears to vary from +2.55 to 2.25m between Burnham and Blakeney

MHW appears to vary from +2.20 to 1.90m

(Only between MHWS-MHW and MHW does the saltmarsh seem to have lower saltmarsh characteristics.)

The most repeated levelling of saltmarsh surfaces on the North Norfolk Coast known to the authors is on Scolt Head Island (Steers, 1960; Allison, 1985; French, 1993; French et al., 1995). It is these OD levels that are most helpful in establishing saltmarsh levels relative to tidal levels and OD values in the Burnham Overy Staithe (Brancaster) area.

MEAN SEA LEVEL (MSL) DURING THE LAST 7,000 YEARS

The rise of global sea-level in the late Devensian and following Holocene, as water from the melting continental glaciers returned to the oceans, is now relatively well-known and understood. Eustatic water adjustments and isostatic adjustments of the coastal crust has produced a drowning of the coastal zone that has continued long after the return of the main glacial ice volume to the oceans. The initially fast rate has passed into a later, slower rate of relative sea-level rise that has been inferred from data obtained, and related to radiocarbon dating in Fenland (nearest to the North Norfolk Coast) and other places around the British Isles. Lambeck (1995) has modelled and contoured the level of Mean Sea Level (MSL) in Fenland and the southern North Sea until 7000 BP, that is essentially relative to present-day OD. The modelled interpretation of the history of sea-level change in and around the British Isles, during late Devensian and Holocene time produced by Lambeck (1995) is very helpful in connection with our investigation on the North Norfolk Coast. It is not necessary to explain the details of that interpretation in the context of this

paper, but only to say that its results have been shown to be consistent with direct evidence of sea-level change deduced from coastal sediments ranging from the Thames estuary to as far as the north of Scotland.

At 7000 BP, MSL off the North Norfolk Coast is shown to have been generally c.21 metres below present-day MSL (effectively present-day OD). Actually MSL at 7000 BP. ranged from c.20.30m below present-day MSL (=c. -20.30m OD) near Burnham Overy Staithe (Brancaster) in the west, through c.-21.00m MSL=OD at Wells, to c.-21.70m MSL=OD. at Blakeney in the east, compared with c.-17.40 m MSL=OD in Fenland (Lambeck 1995, p. 445, fig. 3h). As Lambeck (1995, 441, fig. 2g) also presents the sea-levels he has predicted, (and compares them with values observed by Shennan (1986)), for Fenland) from c.7000 BP until the present day, it is possible to extrapolate and estimate the post-7000 BP history of sea-level change along the North Norfolk Coast, on the reasonable assumption that it proceeded at essentially the same rate as in the Fenland region. In this way a plot of MSL changes can be constructed for the North Norfolk Coast over the last 7,000 years. For the present paper the values for rate of sea-level rise in Table 3 and Figure 1 have simply been measured from Lambeck's (1995, fig. 2g) predicted values as accurately as possible. The differences between the Fenland, Burnham Overy Staithe (Brancaster), Wells and Blakeney Sectors are based on the proportionate difference showing at 7000 BP on Lambeck's (1995) fig.3h. The values in Table 3 are used when calculating ages, when it is assumed that MSL was equivalent to OD.

AGE CALCULATION (ESTIMATION) METHOD

The calculation mechanics

(All results are presented in Appendix 1)

A. The likely minimum (mslmin), i.e. greatest, depth of MSL relative to present-day OD is calculated by taking the OD depth of the sample and adding to it the value above OD that characterises the *upper* tidal limit of the saltmarsh type that has been identified.

Example: Sample from -4.00m OD in borehole NNC40.

Present day depth of -4.0m, plus likely OD equivalent level at time of formation of the +2.9m OD of upper saltmarsh *upper* tidal level, indicates minimum MSL of -6.90m OD at that time.

Table 3. Progress of mean sea-level rise to Ordnance Datum from 7000 BP to the present-day, presented for Fenland, Burnham Overy Staithe (Brancaster), Wells and Blakeney on the N. Norfolk coast.

BP	Fenland	Brancaster	Wells	Blakeney
0	0.00	0.00	0.00	0.00
-500	-0.46	-0.54	-0.56	-0.57
-1000	-0.89	-1.04	-1.07	-1.11
-1500	-1.37	-1.60	-1.65	-1.71
-2000	-1.85	-2.16	-2.23	-2.31
-2500	-2.45	-2.86	-2.96	-3.06
-3000	-3.02	-3.52	-3.64	-3.77
-3500	-3.77	-4.40	-4.55	-4.70
-4000	-4.47	-5.22	-5.39	-5.57
-4500	-5.32	-6.21	-6.42	-6.63
-5000	-6.13	-7.15	-7.40	-7.64
-5500	-7.76	-9.05	-9.37	-9.68
-6000	-9.47	-11.05	-11.43	-11.81
-6500	-12.17	-14.20	-14.69	-15.18
-7000	-17.40	-20.30	-21.00	-21.70

To produce this Table Burnham Overy Staithe (Brancaster), Wells and Blakeney mean sea-levels have *simply been measured* off the Lambeck (1995) map *relative* to one another and to Fenland at the latest time (7000 BP) they are related by Lambeck (1995) to one another, and then *simply calculated pro-rata* at 500 year intervals using *measurements* against the general Fenland mean sea-level rise curve as presented in Lambeck (1995). It is fully realised that the values given cannot be precise or even always realistic, (certainly not to the nearest centimetre!), but they do provide an objective consistency and convenience for the present purpose.

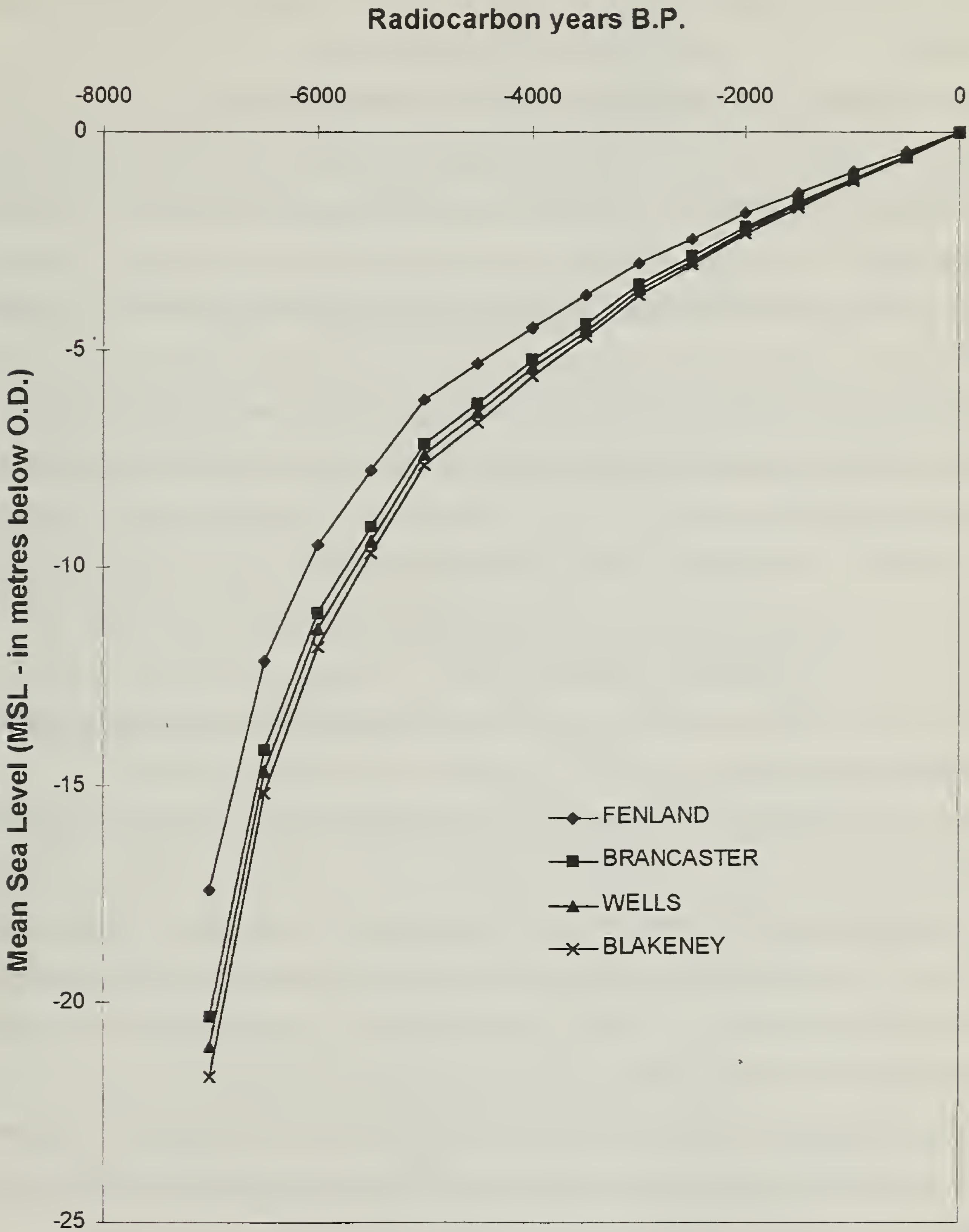

Fig. 1. Progress of sea-level rise from 7,000 B.P. to the present-day, plotted for Fenland, and for Brancaster, Wells and Blakeney on the N. Norfolk coast from data presented in Table 3.

B. The likely maximum (mslmax), i.e. shallowest, depth of MSL relative to present-day OD is calculated by taking the OD depth of the sample and adding to it the value above OD that characterises the *lower* tidal limit of the saltmarsh type that has been identified.

Example: Sample from -4.00m OD in borehole NNC40.

Present day depth of -4.0m, plus likely OD equivalent level at time of formation of the +2.25mOD of upper saltmarsh *lower* tidal level, indicates maximum MSL of -6.25m OD at that time.

C. The age as BP maximum (BPmax) is calculated from the mslmin, i.e. deepest MSL.

Take BP age from Table 3 that is the 500 year younger value than the mslmin indicates. Then calculate the proportionate age distance of the actual mslmin value to the next 500 year value as follows.

Example: Sample from -4.00m OD in borehole NNC40

Calculation is made using the following formula:

4500+((-6.9-6.63)/1.01)*500

where 4500 is the BP age in Table 3 which has a MSL value of -6.63 and is younger than the mslmin value of -6.9;

where 1.01 is the difference in MSL between the 5000 BP value of -7.64 and the 4500 BP value of -6.63.

By determining the MSL difference between the 4500 BP value and the sample value, dividing it by the MSL difference in metres between 4500 BP and the 500 year older age 5000 BP, then multiplying it by 500, a proportionate MSL distance between the younger BP age and the next oldest 500 year stage is converted into a proportionate age.

D. The age of BP minimum (BPmin) is calculated from the mslmax, i.e.. shallowest MSL. Calculation is in all respects, except that it uses mslmax the same as for BPmax above.

E. BPav, BP+/-, AD/BC are simple calculations based on the BPmax and BPmin results.

All ages calculated are presented in Appendix 1. We have also summarised the age determination technique graphically (Fig. 2) using an example of a lower saltmarsh assemblage from a depth of -5.48m OD implying an age of approximately 5ka BP.

VALIDATING THE TECHNIQUE

Although this technique was originally devised as a ranging tool in the absence of other dating methods, we have subsequently obtained a number of radiocarbon dates on organic material from some of the same cores which suggest that the method is surprisingly accurate. In Figure 3 we summarise dates from four Holocene cores in North Norfolk where we now have both biofacies and radiocarbon dates. These indicate that biofacies calculations of depositional ages are consistent with the radiocarbon dates. The offset at the base of cores NNC17 & NNC35, where there are much older peat dates relative to the immediately succeeding sequence, is likely to reflect initially non-marine peat formation compared with the overlying silts, and there may also be an erosional unconformity at the top of these 'basal' peats.

The plots in general give us some confidence in the general technique as a ranging tool in spite of several uncertainties in its construction, and suggests that it may prove useful in similar situations elsewhere.

A number of preliminary comparisons have been made with intercalated radiocarbon dates and some of these are provided in Figure 3. The potential of the method may not appear too great, but it is very generally applicable, it is clearly more, rather than less accurate and precise as it ranges back in time, and it has considerable scope for providing a general time framework for intertidal sedimentary sequences.

Sources of potential error in our estimates

There are clearly several sources of potential error in the basic assumptions which we have made about the relationships made in our estimates between different elements.

1. Assessment of saltmarsh types, that are defined by plant associations, or by using benthic fossil foraminifera (mainly) and ostracods, are subject to inconsistencies, but, for more than two decades investigation of foraminiferal populations in saltmarsh environments have revealed overall consistent associations. As far as the North Norfolk Coast is concerned some Holocene upper saltmarsh deposits contain few or no benthic foraminifera. Those that do contain benthic foraminifera at the present-day are not

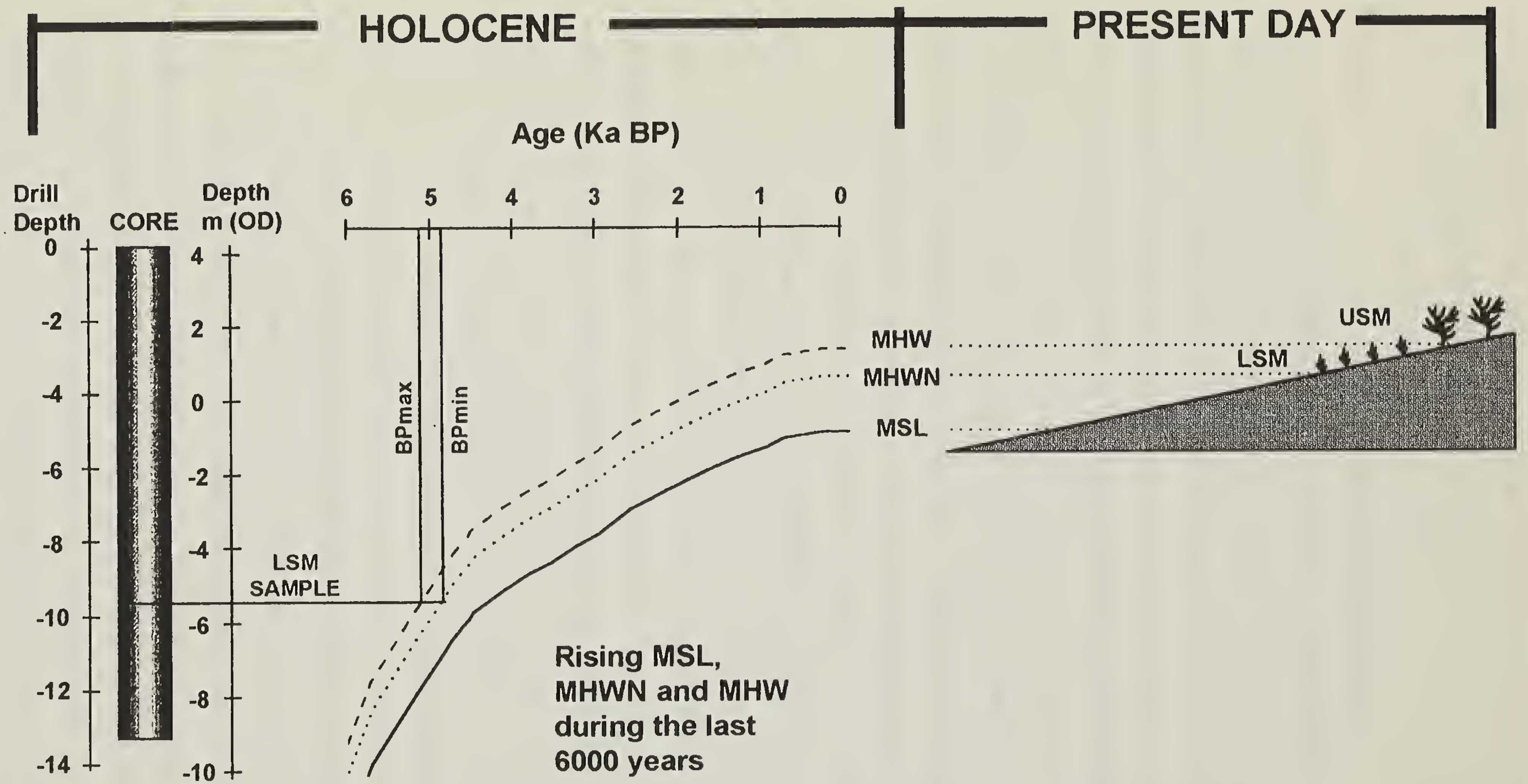

Fig. 2. Schematic representation of the method. On the right hand part of the diagram are the modern relationships between tidal levels (MHW, MHWN and MSL) and intertidal environments (lower saltmarsh, LSM and upper saltmarsh, USM). The middle part of the diagram shows the rate of mean sea-level rise for the area (after Lambeck, 1995) with commensurate changes in MHWN and MHW during the same period. The left hand part of the diagram illustrates the worked example in the text with a LSM sample taken from a core depth of 9.50m. This depth equates to an elevation of -5.48m O.D. implying an age between 4910 and 5139 years B.P.

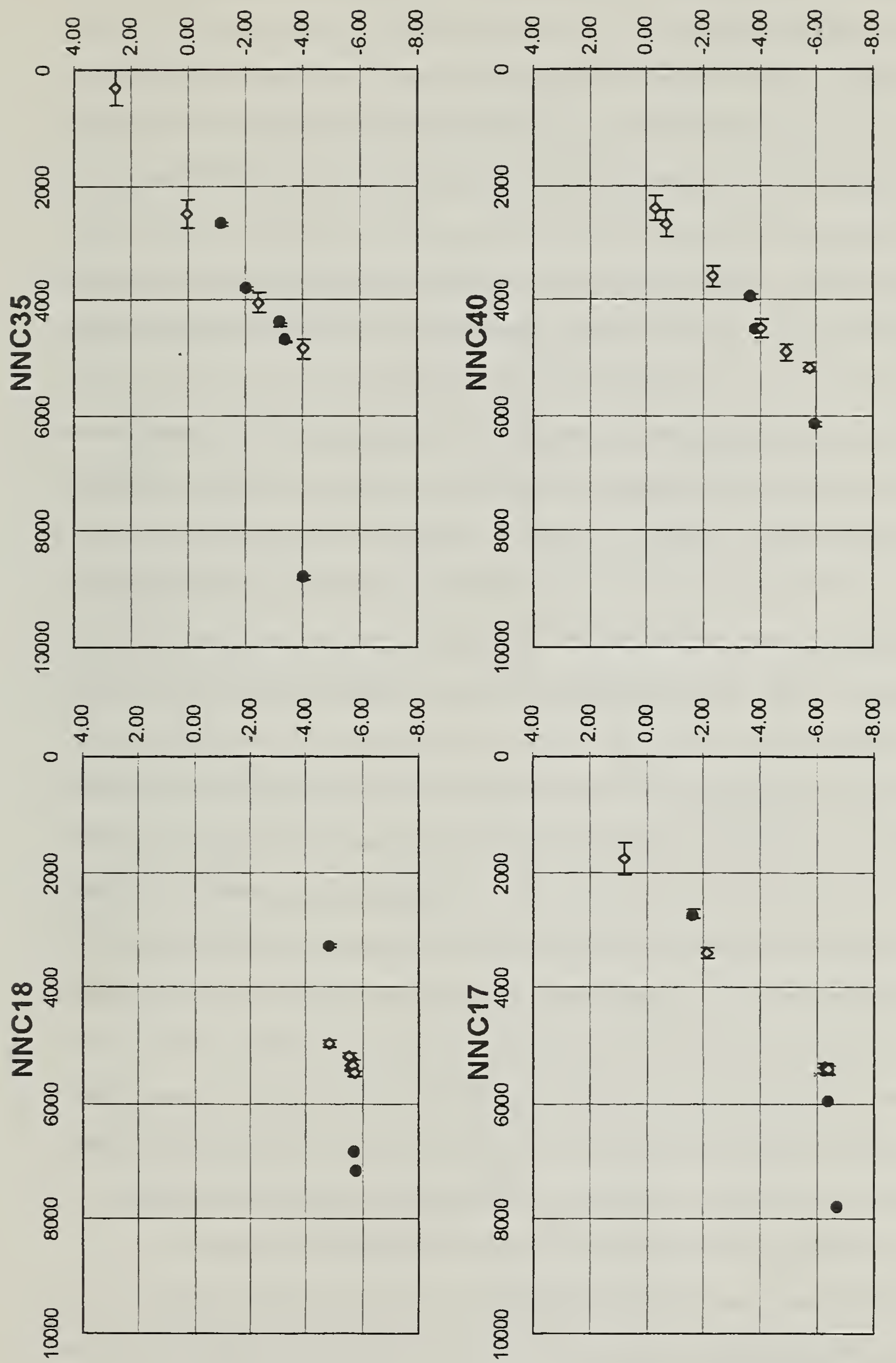

Fig. 3. Age (years B.P.) versus depth (metres, relative to O.D.) plots for four cores through the Holocene of North Norfolk. Closed circles are radiocarbon ages (average error ±80 years), open diamonds are biofacies determined ages.

accumulating above HAT-MHWS, so we have distinguished upper-upper saltmarsh associations (uusm = no foraminifera, *usm = very few foraminifera) in the Holocene record in Appendix 1, and make the assumption for the present purpose that they accumulated above HAT-MHWS (but obviously below HAT).

HAT-MHWS to MHWS-MHW have been assumed to define the upper saltmarsh assemblages, using OD values for these tidal limits obtained from the Admiralty Tidal Tables, (and mainly from measurements and observations on Scolt Head Island), for calculations in the present paper;

MHWS-MHW to MHW have been assumed to define the lower saltmarsh assemblages, again using OD values for these tidal limits from the Admiralty Tidal Tables, (and on Scolt head island), for calculations in the present paper.

[The appreciation of microfossil intertidal environments in terms of tidal ranges, quoted earlier in the paper, placed the uppermost saltmarsh/upper saltmarsh boundary at MHWS, and the upper saltmarsh/lower saltmarsh boundary at MHW. These tidal levels do not correspond very well with the OD levels best known for upper and lower saltmarsh on Scolt Head Island. We have therefore defined the expected microfossil upper and lower saltmarsh biofacies in terms of the OD levels, (*and the corresponding tidal levels*), that they appear to use.]

2. Because of the higher rate of sea-level rise at that time, the earlier Holocene estimates appear to be more precise and accurate than those for the later Holocene. This is the opposite to the trend in precision and accuracy to be expected from radiocarbon dates. The Holocene sea level reconstruction we have used has errors which may not be inconsiderable; Lambeck (1995) suggests these may be as much as 1-2m for the mid-late Holocene.

3. Sediment compaction is also likely to affect the accuracy. It seems likely, however, that the most significant sediment compaction processes affect only the topmost layers of recently accumulated saltmarsh sediment, and that the continuation of compaction in more deeply buried sediments is much more limited (Pye, 1992).

Having admitted these sources of potential error, our initial results suggest that we have developed a useful ranging technique which provides useable, consistent, and not altogether inaccurate age estimates, which compare rather well with available radiocarbon dates.

ACKNOWLEDGEMENTS

Almost all the results reported in this paper have been produced from research undertaken using support provided as part of the Natural Environment Research Council's (NERC's), Community Research Project: Land-Ocean Interaction Study (LOIS): Land Ocean Evolution Perspective Study (LOEPS) Component. These particular results were derived from Special Topic No. 32; GST/02/737 "Sedimentary Evolution of the North Norfolk Barrier Island Coastline in the context of Holocene Sea Level Change", awarded to Dr. G.M. Harwood (School of Environmental Sciences, University of East Anglia) and others (1994-1997). (Since Dr. Harwood's tragic death in the Spring of 1996 this special topic project has been overseen by Dr. J.E. Andrews.)

Sedimentary, stratigraphical and micropalaeontological data has been compiled by Drs. R.W. Jones, I. Boomer, and Prof. B.M. Funnell at UEA, particularly collaborating with Dr. C.S. Bristow of Birkbeck College in relation to the western sequences that include Scolt Head Island.

Geophysical, palaeomagnetic, luminescence and radiometric results have been, or are being obtained, either as part of the UEA programme (Dr. P.N. Chroston and Dr. R. Jones - geophysics, Dr. B.A. Maher and Dr. I. Boomer - palaeomagnetics), or as part of the Durham University (Mr. I.K. Bailiff and Dr. H. Roberts- luminescence) or Edinburgh University (Dr. G.B. Shimmield and Mr J. Smith - radionuclide data; GBS is now at Dunstaffnage Marine Laboratory).

We are grateful to Prof. Ian Shennan, and his collaborators at Durham University for results and permission to use results, (produced by Mr. J. Innes, on pollen; Dr. J. Lloyd and Dr. W.R. Gehrels, on benthic foraminifera), from peat and associated clay samples taken from LOEPS boreholes at levels adjacent to radiocarbon samples.

This is LOIS publication number 311.

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APPENDIX 1

Biofacies age estimates for all micropalaeontologically examined saltmarsh samples obtained from LOEPS boreholes on the North Norfolk Coast.

For the *calculation* of age estimates:

In the Burnham (Brancaster) Sector

it is assumed that:	+3.50 to +3.20m OD =*usm and uusm
	+3.20 to +2.55m OD = usm
	+2.55 to +2.20m OD = lsm
In the Wells Sector it is assumed that:	+3.05 to +2.40m OD = usm
	+2.40 to +2.05m OD = lsm
In the Blakeney Sector it is assumed that:	+3.20 to +2.90m OD = uusm
	+2.90 to +2.25m OD = usm
	+2.25 to +1.90m OD = lsm

The abbreviations used are as follows:

mOD	=	sample level relative to OD
env	=	inferred environment
dur\	=	micropalaeontological data provided by Durham University (all other micropalaeontological data obtained by Ian Boomer)
uusm	=	upper-upper saltmarsh, possibly even brackish or freshwater = (MHWS-HAT, or higher than HAT)
*usm	=	upper-upper saltmarsh, limited marine microfauna = (MHWS-HAT)
usm	=	upper saltmarsh = (MHW-MHWS)
lsm	=	lower saltmarsh = (MHWN-MHW)
e/	=	euryhaline
b/	=	brackish
mslmin	=	inferred minimum Mean Sea Level
mslmax	=	inferred maximum Mean Sea Level
BPmax	=	Estimated maximum Age in years Before Present
BPmin	=	Estimated minimum Age in years Before Present
BPav	=	Estimated mean Age in years Before Present
BP±	=	Range of Age estimate relative to mean Age Before Present
AD/BC	=	Age AD or (-)BC

Burnham (Brancaster) Sector

Thornham Section

NNC34 (1.99m OD)* [Nat. Grid. Ref. TF 72170 44410]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
-0.31	b/usm	-3.51	-2.86	2992	2500	2746	246	-796
-0.91	usm	-4.11	-3.46	3335	2955	3145	190	-1195

NNC35 (2.75m OD)* [Nat. Grid. Ref. TF 73504 44641]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
2.55	usm	-0.65	0.00	610	0	305	305	1645
0.05	usm	-3.15	-2.50	2720	2243	2481	238	-531
-2.45	usm	-5.65	-5.00	4217	3866	4042	176	-2092
-3.97	usm	-7.17	-6.52	5005	4665	4835	170	-2885

Brancaster - Scolt Head Island Section

NNC29 (7.07m OD) [Nat. Grid. Ref. TF 78290 45400]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
-2.10	usm	-5.30	-4.65	4040	3652	3846	194	-1896
-3.63	usm	-6.83	-6.18	4830	4485	4657	172	-2707

NNC30 (4.55m OD) [Nat. Grid. Ref. TF 77490 45220]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
1.15	usm	-2.05	-1.40	1902	1321	1612	290	338

NNC21 (4.13m OD) [Nat. Grid. Ref. TF 83514 46248]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
1.43	usm	-1.77	-1.12	1652	1071	1362	290	588
0.88	e/lsm	-1.67	-1.32	1563	1250	1406	156	544

Burnham Section

NNC18 (4.02m OD) [Nat. Grid. Ref. TF 85765 45898]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
-4.78	e/lsm	-7.33	-6.98	5047	4910	4978	69	-3028
-5.48	lsm	-8.03	-7.68	5232	5139	5186	46	-3236
-5.57	dur\usm	-8.77	-8.12	5426	5255	5341	86	-3391
-5.60	usm	-8.80	-8.15	5434	5263	5349	86	-3399
-5.62	dur\usm	-8.82	-8.17	5439	5268	5354	86	-3404
-5.66	dur\usm	-8.86	-8.21	5450	5279	5364	86	-3414
-5.68	dur\uusm	-9.18	-8.88	5533	5458	5495	38	-3545

NNC19A (3.07m OD) [Nat. Grid. Ref. TF 85814 45644]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
-1.09	usm	-4.29	-3.64	3438	3068	3253	185	-1303
-3.65	e/lsm	-6.20	-5.85	4495	4318	4407	88	-2457
-3.93	usm	-7.13	-6.48	4989	4644	4816	173	-2866
-4.24	e/lsm	-6.79	-6.44	4809	4622	4715	93	-2765
-6.15	dur\lsm	-8.70	-8.35	5408	5316	5362	46	-3412

NNC20 (2.17m OD) [Nat. Grid. Ref. TF 85450 45090]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
0.37	usm	-2.83	-2.18	2479	2014	2246	232	-296
-0.18	usm	-3.38	-2.73	2894	2407	2651	243	-701
-0.48	usm	-3.68	-3.03	3091	2629	2860	231	-910

NNC20A (2.51m OD) [Nat. Grid. Ref. TF 85500 45080]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
2.06	usm	-1.14	-0.49	1089	454	771	318	1179
1.76	usm	-1.44	-0.79	1357	750	1054	304	896
1.31	usm	-1.89	-1.24	1759	1179	1469	290	481
1.01	usm	-2.19	-1.54	2021	1446	1734	288	216

NNC23 (2.85m OD) [Nat. Grid. Ref. TF 83440 45160]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
1.45	*usm	-2.05	-1.75	1902	1634	1768	134	182
0.50	*usm	-3.00	-2.70	2606	2386	2496	110	-546
-0.45	*usm	-3.95	-3.65	3244	3074	3159	85	-1209
-1.05	*usm	-4.55	-4.25	3591	3415	3503	88	-1553

NNC24 (2.36m OD) [Nat. Grid. Ref. TF 83360 44160]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
1.76	*usm	-1.74	-1.44	1625	1357	1491	134	459
0.76	*usm	-2.74	-2.44	2414	2200	2307	107	-357
0.30	*usm	-3.20	-2.90	2758	2530	2644	114	-694

Wells Sector

Holkham Section

NNC16 (4.02m OD) [Nat. Grid. Ref. TF 89030 4500]

mOD	env	mslmin	mslmax	BPmax	BPmin	BPav	BP±	AD/BC
-6.78	usm	-9.83	-9.18	5612	5452	5532	80	-3582
-6.98	e/lsm	-9.38	-9.03	5502	5414	5458	44	-3508
-7.94	dur\lsm	-10.34	-9.99	5735	5650	5693	42	-3743
-7.98	dur\lsm	-10.38	-10.03	5745	5660	5703	42	-3753
-8.00	dur\lsm	-10.40	-10.05	5750	5665	5708	42	-3758

NNC17 (2.37m OD) [Nat. Grid. Ref. TF 89130 44550]

mOD	env	mslmin	mslmax	BPmax	BPmin	BPav	BP±	AD/BC
0.77	usm	-2.28	-1.63	2034	1483	1759	276	191
-2.08	e/lsm	-4.48	-4.13	3462	3269	3365	96	-1415
-6.24	dur\usm	-9.29	-8.64	5480	5315	5397	82	-3447
-6.28	dur\usm	-9.33	-8.68	5490	5325	5407	82	-3457
-6.30	dur\usm	-9.35	-8.70	5495	5330	5412	82	-3462
-6.32	dur\usm	-9.37	-8.72	5500	5335	5418	82	-3468
-6.34	dur\usm	-9.39	-8.74	5505	5340	5423	82	-3473

Wells Section

NNC14 (3.10m OD) [Nat. Grid. Ref. TF 94840 44300]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD BC*
2.90	usm	-0.15	0.50	134	-446	-156	290	2106
-0.80	usm	-3.85	-3.20	3115	2676	2896	219	-946
-2.80	usm	-5.85	-5.20	4223	3887	4055	168	-2105
-3.80	usm	-6.85	-6.20	4719	4393	4556	163	-2606
-5.10	usm	-8.15	-7.50	5190	5025	5108	82	-3158
-5.86	usm	-8.91	-8.26	5383	5218	5301	82	-3351
-6.60	usm	-9.65	-9.00	5568	5406	5487	81	-3537
-7.80	e/lsm	-10.20	-9.85	5701	5617	5659	42	-3709
-8.80	e/lsm	-11.20	-10.85	5944	5859	5902	42	-3952
-10.00	e/lsm	-12.40	-12.05	6149	6095	6122	27	-4172
-10.60	e/lsm	-13.00	-12.65	6241	6187	6214	27	-4264

Blakeney Sector

Blakeney Section

NNC03 (3.37m OD) [Nat. Grid. Ref. TF 99816 85894]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD BC*
-7.70	lsm	-9.95	-9.60	5563	5480	5522	41	-3572

NNC04 (2.96m OD) [Nat. Grid. Ref. TG 00965 46418]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD BC*
-11.49	dur\uusm	-14.69	-14.39	6427	6383	6405	22	-4455
-11.53	dur\uusm	-14.73	-14.43	6433	6389	6411	22	-4461
-11.59	dur\uusm	-14.79	-14.49	6442	6398	6420	22	-4470
-11.65	dur\uusm	-14.85	-14.55	6451	6407	6429	22	-4479

Cley Section

NNC01 (1.85m OD) [Nat. Grid. Ref. TG 04910 44620]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
0.75	usm	-2.15	-1.50	1867	1325	1596	271	354
0.35	usm	-2.55	-1.90	2160	1658	1909	251	41
-0.40	usm	-3.30	-2.65	2669	2227	2448	221	-498
-1.00	usm	-3.90	-3.25	3070	2634	2852	218	-902
-1.38	b/usm	-4.28	-3.63	3274	2901	3088	186	-1138
-4.92	b/lsm	-7.17	-6.82	4767	4594	4681	87	-2731
-5.25	b/lsm	-7.50	-7.15	4931	4757	4844	87	-2894
-8.02	e/usm	-10.92	-10.27	5791	5638	5715	76	-3765

NNC40 (1.90m OD) [Nat. Grid. Ref. TG 08070 44000]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD/BC*
-0.30	usm	-3.20	-2.55	2599	2160	2379	219	-429
-0.70	usm	-3.60	-2.95	2880	2427	2653	227	-703
-2.30	usm	-5.20	-4.55	3787	3419	3603	184	-1653
-4.00	usm	-6.90	-6.25	4634	4330	4482	152	-2532
-4.90	usm	-7.80	-7.15	5039	4757	4898	141	-2948
-5.70	usm	-8.60	-7.95	5235	5076	5156	80	-3206

Morston Section

83-11 (2.30m OD) [Nat. Grid. Ref. TF 99950 44750]

mOD	*env*	*mslmin*	*mslmax*	*BPmax*	*BPmin*	*BPav*	*BP±*	*AD BC*
1.70	usm	-1.20	-0.55	1075	482	779	296	1171
1.00	usm	-1.90	-1.25	1658	1117	1388	271	563
0.00	usm	-2.90	-2.25	2393	1950	2172	222	-222
-1.00	usm	-3.90	-3.25	3070	2634	2852	218	-902
-2.00	usm	-4.90	-4.25	3615	3258	3437	178	-1487
-3.00	usm	-5.90	-5.25	4160	3816	3988	172	-2038
-4.00	usm	-6.90	-6.25	4634	4330	4482	152	-2532

CONTENTS

The Geological Society of Norfolk exists to promote the study and understanding of geology, particularly in East Anglia, and holds monthly meetings throughout the year.

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The illustration on the front cover is part of figure 2 from the article by Funnell and Boomer in this issue of the Bulletin.

ISSN0143-9286

BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

(FOR ARTICLES ON THE GEOLOGY OF EAST ANGLIA)

NO.47 1997

PUBLISHED 1998

CONTENTS INCLUDE

thematic set of papers
on
Belemnitella
in the Beeston Chalk of Norfolk

BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

No. 47 (for 1997) Published 1998

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EDITORIAL

Bulletin No. 47 comprises three related papers by Mike Godwin, all dealing with aspects of the belemnite palaeontology and palaeoecology of the Beeston Chalk in Caistor St. Edmunds quarry near Norwich, Norfolk. The first paper discusses preliminary data on growth rings in *Belemnitella* guards that suggest life spans of 200-300 days. The second paper describes a possible occurrence of *Belemnitella cf. americana,* while the third paper is a broader review of palaeoenvironments and cyclicity of the Beeston Chalk, showing how these might link with the palaeoecology of *Belemnitella*.

The appearance of Bulletin 47 brings the publication schedule up to date. I will be assembling copy for Bulletin 48 during the year and welcome the submission of papers on any aspect of East Anglian geology. I hope to publish Bulletin 48 during the winter of 1998/99 to keep the publication schedule up to date. I would like to thank all the authors and reviewers who have contributed their work and time to Bulletins 42 to 47 and I look forward to assembling future editions.

INSTRUCTIONS TO AUTHORS

If possible, contributors should submit manuscripts as word-processor print out accompanied by a disk copy. We can handle most word-processing formats although PC Word, WordPerfect or ASCII files are preferred. In addition we accept typewritten copy and will consider legible handwritten material.

It is important that the style of the paper, in terms of overall format, capitalisation, punctuation, etc. conforms as strictly as possible to that used in Vol. 41 of the Bulletin. Titles and first order headings should be capitalised, centred and in bold print. Second order headings should be centred, bold and lower case. Text should be 1½ line spaced. All measurements should be given in metric units.

References should be arranged alphabetically in the following style.

BALSON, P.S. & CAMERON, T.T.J. 1985. Quaternary mapping offshore East Anglia. *Modern Geology*, **9**, 221-239.

STEERS, J.A. 1960. Physiography and evolution: the physiography and evolution of Scolt Head Island. In: Steers, J.D. (ed.) *Scolt Head Island* (2nd ed.), 12-66, Heffer, Cambridge.

BLACK, R.M. 1988. *The Elements of Palaeontology*. 2nd Ed., Cambridge University Press, Cambridge. 404pp.

Illustrations should be drawn with thin dense black ink lines. Thick lines, close stipple or patches of solid black should be avoided as these spread in printing. Original illustrations should, before reproduction, be not more than **175mm by 255mm**. Full use should be made of the first (horizontal) dimension which corresponds to the width of print on the page, but the second (vertical) dimension is an upper limit only. Half tone photographic plates are acceptable when their use is warranted by the subject matter, provided the originals exhibit good contrast.

The editors welcome original research papers, notes or comments, and review articles relevant to the geology of **East Anglia** as a whole, and do not restrict consideration to articles covering Norfolk alone. All papers are independently refereed by at least one reviewer.

GROWTH RINGS AND POPULATION DYNAMICS OF *BELEMNITELLA* IN THE BEESTON CHALK, UPPER CAMPANIAN, CAISTOR ST. EDMUND, NORFOLK.

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ABSTRACT

*The **Belemnitella** populations in a section of the Beeston Chalk (Upper Campanian) are examined in terms of their diameter at the protoconch. An attempt is made to determine the relative ages of a number of specimens by counts and estimates of the total number of growth rings. The current views of belemnite palaeobiology are reviewed and revised in the light of recent advances in cephalopod research. The analysis of growth rings on a number of specimens leads to a tentative conclusion that species of **Belemnitella** had a life-span of c. 200-300 days.*

*Analysis of populations within the Beeston Chalk show that the **B. mucronata** group is dominant on hardgrounds and that the **B. langei** group dominates in the soft white chalks. The **B. mucronata** group is only represented by adult or near adult individuals whilst the **B. langei** group is represented by juveniles and adult forms. This suggests the latter group was endemic to the Caister area and that the **mucronata** group migrated there from some distance away. The observed variations suggest cyclicity, possibly related to sea level change.*

INTRODUCTION

In this paper an attempt is made to determine the life spans of Upper Cretaceous belemnites by counting growth rings within their guards. This study should be regarded as a preliminary investigation and further work on more specimens from a greater range of

Bull. geol. Soc. Norfolk (for 1997) 47, 3-18 (Published 1998)

species is clearly desirable. The relative age ranges and the population distributions of the two main groups of species are also examined through a short section of Upper Campanian chalk.

The specimens described here come from the Beeston Chalk Member (Upper Campanian of the Chalk of Norfolk) from the quarry at Caister St. Edmunds (Grid Reference TG 2390 0466). The lithostratigraphy and biostratigraphy of the Campanian chalks of Norfolk have been fully discussed in the past by Jeletzky (1951), Peake and Hancock (1970), Wood (1988), Johansen and Surlyk (1990), Pitchford (1991a; 1991b) and Christensen (1995) and are not discussed in detail here. The sequence of beds at Caistor St. Edmund is given in Figure 1. The beds are located in the *B. minor* I zone (Christensen, 1995).

The strata selected for study, consists of approximately 4.6m of phosphatic chalk, and occurs from just below Flint Band 7 to Flint Band 12 of Wood (1988; see Fig. 1). This includes the *Echinocorys* Bed. Flint Band 7 is equivalent to Flint Band Z of Peake and Hancock (1970) and Flint Band 3 of Pitchford (1991a).

The lowest levels of the exposure, including the hardground below Flint Band 7, are now filled in and no longer accessible. Recently a large volume of chalk has been removed from the pit. This has rendered the exposure between Flint Bands 7 and 10 inaccessible or dangerous.

The greatest concentration of belemnites is found on the incipient hardground below Flint Band 7. This was not noted by Wood (1988), as it was not well exposed at that time but it was observed by Pitchford (1991a), and he described it as "intermittent". Wood, however, observed a greater concentration of belemnites just above Flint Band 5 on what he interpreted to be an omission surface. Belemnites are also common in the *Echinocorys* Bed and very abundant between Flint Bands 10 and 11. Pitchford (1989) noted other horizons within the Beeston Chalk with abundant belemnites but these were not encountered in the present study.

MATERIALS

These observations are based on a collection of over 300 guards, many of which are fragmentary. Following recent taxonomic analysis of *Belemnitella* in the Beeston Chalk (Christensen 1995), five species have been identified from a few good specimens: *Belemnitella minor* I Jeletzky; *Belemnitella langei* Jeletzky; *Belemnitella sp.* 1

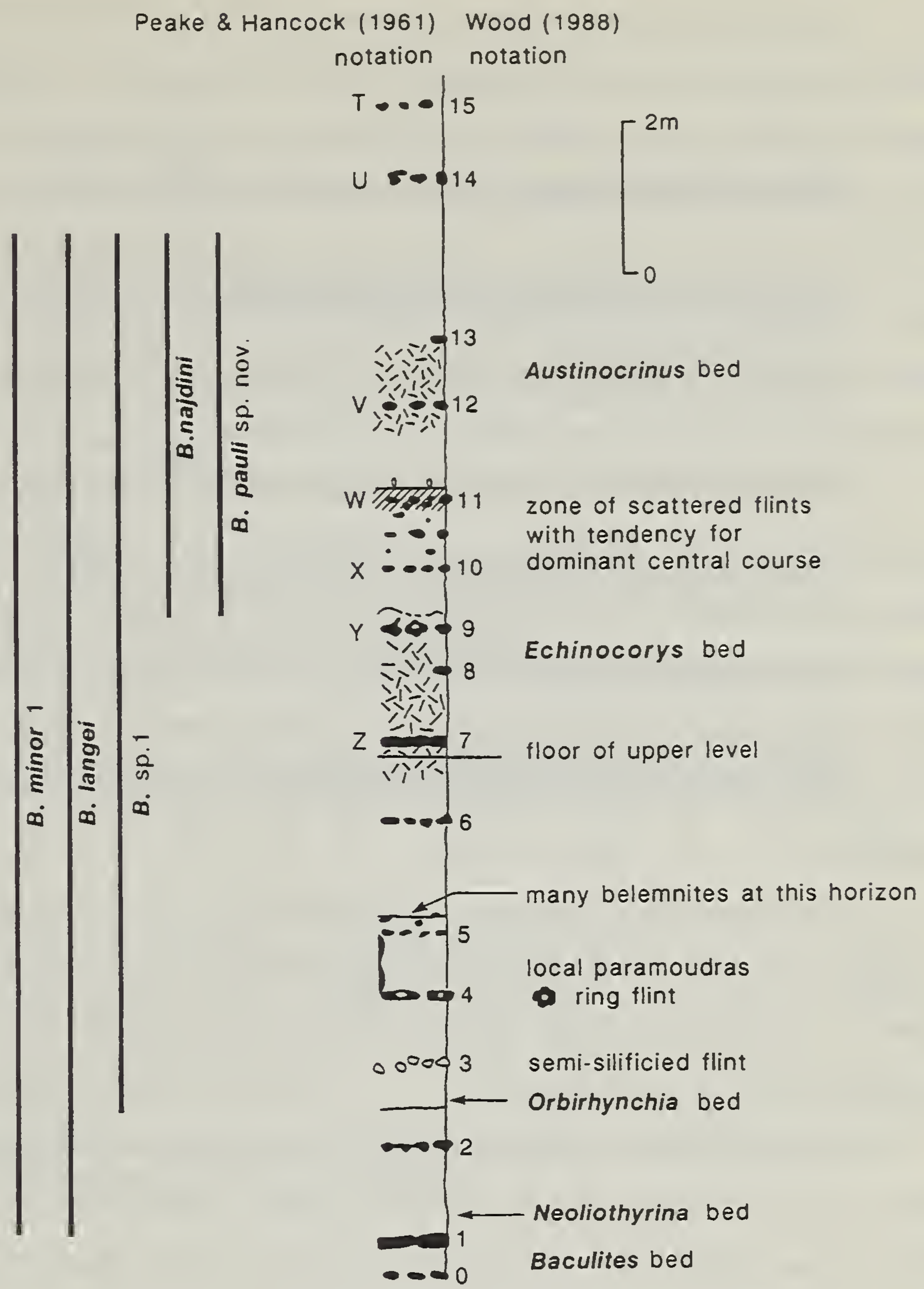


Fig. 1. The stratigraphy of the Beeston Chalk at Caistor St. Edmund and the ranges of *Belemnitella* sp., after Wood (1988) and Christensen (1995).

Christensen; *Belemnitella najdini* Kongiel; and *Belemnitella pauli* sp. nov. Christensen. *B*. sp.1 appears to be similar to *B. mucronata postrema* Naidin.

Morphometric analysis was carried out on 63 specimens, from the hardground below Flint Band 7; 61 specimens, from the *Echinocorys* Bed; 97 specimens, from between Flint Bands 10 and 11; and 20 specimens from the hardground above Flint Band 11 in which the alveolus was preserved in its entirety or was near complete. The collection was made over a period of six years but the majority of specimens were recovered in April 1995 (Flint Bands 7-9) and May 1996 (Flint Bands 9-12).

MORPHOMETRIC ANALYSIS AND POPULATION DYNAMICS

It was not possible (due to the fragmentary nature of many guards) to identify most of the guards to specific level as defined by Christensen (1995). However, most specimens (other than the numerous small fragments) could be assigned to either the *Belemnitella mucronata* Group (*B. minor*, *B. pauli*, *B.* sp.1) or the *Belemnitella langei* group (*B. langei*, *B. najdini*) on the basis of their internal and external morphology and patterns of vascular markings. The populations for each bed or hardground are summarised in Figure 2 in terms of percentage of population and total numbers.

Some entire specimens were split to determine the measurement ratios and angles currently employed in identification (see Christensen, 1975; 1995). These include the dorso-ventral diameter at the protoconch; the length from apex to protoconch (LAP); length from protoconch to base of ventral fissure (Schatzky Distance - SD); alveolar angle (AA); and the fissure angle (FA). These were then compared to the range given by Christensen (1995) for the species from the Upper Campanian in a series of tables and scatter plots.

All five species of *Belemnitella* occur in this part of the Beeston Chalk. However, examples of *B. pauli* and *B. najdini* were not found below Flint Band 10 (as shown in Figure 1). This was probably due to a small total population and a lack of good specimens.

The concentration of belemnites guards on the incipient hardground below Flint Band 7 was dominated by the *B. mucronata* group. (This is also true of the omission surface above Flint Band 9 and the hardground above Flint Band 11). In the chalks between these horizons the *B. langei* group is dominant but the total populations are far smaller in size, with the exception of the chalks lying between Flint Bands 10 and 11 (see Fig. 2).

Belemnite life spans

If the current palaeoecological model of the belemnite life cycle is accepted, that is, that it was similar to modern day coleoids (squid), then the animal would have lived for 1 to 5 years. At this point the animals would have spawned at a localised breeding ground. This would have led to mass mortality among the exhausted participants as much of their body mass would have been converted into the sperm and eggs which sired subsequent generations. This was the most popular theory advanced to explain accumulations of

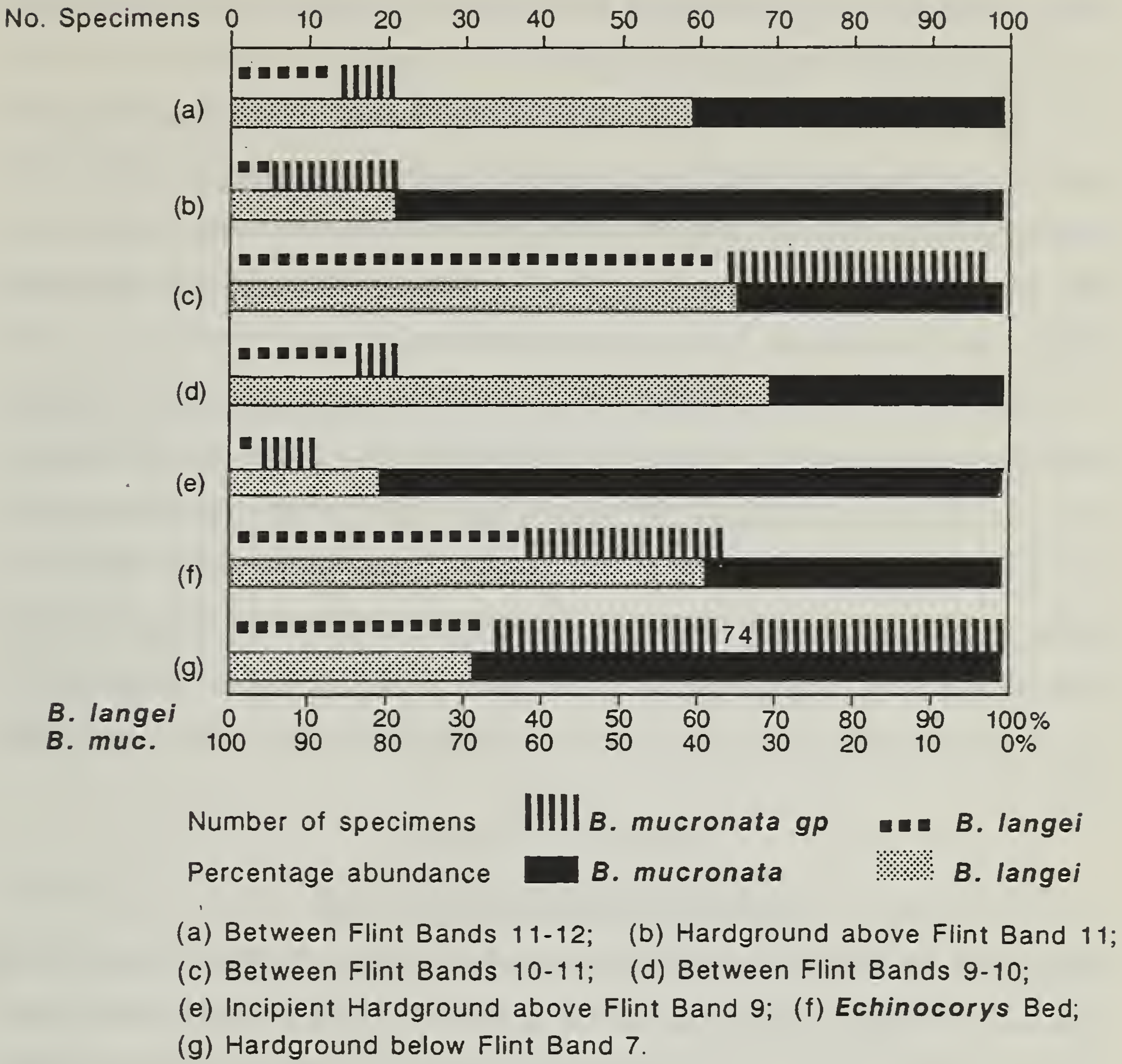


Fig. 2. The relative abundance of the *Belemnitella mucronata* and *langei* groups between Flint Bands 7 and 12 (of Wood, 1988) in the Beeston Chalk.

belemnite guards in the past (e.g. Jarvis, 1980). However, guards may have become concentrated by a variety of causes (see Doyle and Macdonald, 1993). The life-span estimate is based on the interpretation of oxygen isotope analyses of palaeotemperatures derived from belemnite guards (Bowen, 1966; Zhirmunskiy *et al.*, 1967). However, this needs re-examination in the light of recent research, which has provided a better understanding of growth patterns of modern cephalopods.

The life span of oceanic squid is estimated at 3 to 5 years maximum for giant genera such as Architeuthis, but few live longer than 2 years (Clarke, 1966; Packard,

1972; Forsythe and Van Heukelen, 1987). Smaller species attain an adult size very quickly and then growth slows for their brief adult life. The majority lay eggs at the age of 1 to 2 years, and then die (Mangold, 1987).

Recent estimates for full maturation in small coleoids are based on daily growth laminations on statoliths - small, paired, aragonitic head stones (Lipinski, 1986). These stones appear to be common to most cephalopods and are possibly related to the animal's ability to sense linear and angular acceleration and orientation (Clarke, 1978; Williamson, 1995).

Estimates of life spans include: *Loligo chinensis*, *Loliolus noctiluca*, *Idiosepius pygmaeus* (tropical species) - <200 days (Jackson and Choat, 1992); *Loligo vulgaris* - 240 days (Natsukari and Komine, 1992); *Loligo gahi* - 325-345 days (females) and 250-290 days (males) (Arkhipkin, 1993); *Loligo forbesi* - 241-370 (females) and 236-390 (Males) (Collins *et al.*, 1995a). Short life spans of a year or less are thought to be correct by most authors and the maturation rate appears to be temperature controlled with tropical species maturing more rapidly than their temperate cousins (Jackson and Choat, 1992).

Sexual dimorphism and growth rates

The final size of both adult male and female cephalopods may vary greatly within a species depending on food availability and sea temperatures and even siblings reared under laboratory conditions may vary considerably (Forsythe and Van Heukelen, 1987). Sexual dimorphism is common among molluscs but has not been noted as yet in *Belemnitella*, but is known to occur in other belemnite genera (Doyle, 1985). Males of the species *Loligo forbesi* tend to be larger than the females, and the largest are thought to live longer than one year. However, 40% of the population matures at a small size due to different growth rates (Collins *et al.*, 1995). It is likely that the variations in size, growth rates and internal morphology of the various species of *Belemnitella* may be partly caused by sexual dimorphism. However, this remains unquantified at present, and requires extensive bivariate morphological analysis of a large number of specimens.

Growth in *Idiosepius pygmaeus* is thought to be allometric (Lewis and Choat, 1992). Belemnite growth measured from guards is thought to be allometric as well (Christensen, 1975).

Given an abundant food supply, the main factor affecting growth rates is temperature. This in turn is controlled by seasonality, latitude and water depth. These factors are important, considering that individuals of the genus *Belemnitella* may well have had migratory habits through their life cycle, which could include vertical as well as horizontal components.

High temperatures tend to inhibit growth, as metabolic costs increase as animals become more active. Low temperatures decrease the metabolic rate to the point where feeding and growth stop. Cephalopods at the lower temperature limits of their geographical range tend to reach a greater size than their warm-water cousins (Forsythe and Van Heukelen, 1987). This could explain the "giantism" seen in individuals from the chalk between Flint Bands 10 and 12 and in some other horizons.

Belemnites in the Upper Cretaceous appear to have been stenothermal (Bowen, 1966) and therefore restricted to epi-pelagic sea over the continental shelves. Oceanic waters would have been cooler and more variable in temperature (Schonfeld and Burnett, 1991).

Analysis of growth rings

Growth rings appear to have been produced within guards, on a continual basis, as an individual grew. Oxygen isotope analyses of the various growth stages of guards reveal up to five temperature highs or lows. These are assumed to represent seasonality, possibly winter/summer growth pattern (Zhirmunskiy *et al.*, 1967; Urey *et al.*, 1951; Saelen, 1989). Stevens (1965) proposed a 3 to 4 year life cycle. On the other hand, Packard (1972) has observed that coleoids have a phenomenal growth rate and, except for the largest species, a short life-span. As discussed above, recent studies of coleoid growth (Lipinski, 1986; Natsukari and Komine, 1992; Jackson, 1992; Jackson and Choat, 1992; Collins *et al.*, 1995a) demonstrate a life span of 300-400 days for small to medium-sized species in temperate waters. Tropical species have shorter life spans of <200 days (Jackson and Choat, 1992).

Longinelli (1969) and Spaeth (1975) suggest that the observed oxygen isotope temperature variations are secondary diagenetic effects caused by the deposition of diagenetic calcite within pore spaces after death and therefore such results are not reliable. Lowenstam and Epstein (1954) also observed "seasonal" variation in the cuttlebone of modern *Sepia*. This was thought to confirm that belemnites lived for several years due to

the similarity of the temperature patterns observed in modern and fossil material. It is now known that *Sepia*'s life-span is a year or less (Mangold, 1987).

On this basis it is possible to put forward the hypothesis that belemnites similarly had a life-span of 200-500 days. If this was the case, the growth rings present within the guards would represent daily growth increments (as has been found in squid statoliths). It was decided to test this possibility by counting growth rings on a number of specimens.

The dorsal-ventral diameter at the protoconch (DVDP) was used as a measure of relative age for all specimens. It is only in this plane that all the growth stages of a guard are recorded. The relationship between guard diameter and number of growth rings present appeared to be roughly consistent in all specimens of the *B. langei* and *B. mucronata* groups where these could be counted up to a maximum diameter of 20mm at the DVDP. Most guards show some signs of partial or total recrystalisation which obscure part or all of the growth ring sequence. The partial counts were made on 6 specimens from the *Echinocorys* Bed and 1 "giant" (DVDP 20mm) specimen from between Flint Bands 9 and 10 (see Table 1). These belemnites have been identified to specific level.

Growth rings were counted (under a binocular microscope) in 0.5mm segments from the protoconch outwards; at the measured centre of the growth ring sequence; and from the outer edge of the guard inwards. An estimate of the total number of rings in the sequence was obtained by averaging the rate of growth ring accumulation times distance in between partial counts. On one specimen (DVDP 10.2mm) where a total count was possible this revealed an error of - 2.2% between the estimated and actual number of growth rings and therefore the estimates should be treated with caution.

Discussion of results

The results (Table 1) imply that both species of *Belemnitella* had a fast initial growth rate which tapered off as maximum size was reached. This is similar to growth patterns in modern coleoids (e.g. Jackson and Choat, 1992). These authors also define 3 stages in the cephalopod life cycle. Juveniles are 0-100 days old and adulthood begins after around 60 days when sexual maturation (and dimorphism) begins. Full sexual maturity (when eggs or sperm are produced) is reached shortly before death. Juveniles tend to be up to a third of the maximum adult size so in this study *B. langei*'s juveniles are arbitrarily defined as such when their DVDP is 4mm or less and *B. minor*'s at 7mm or less. No juveniles of

Table 1. DVDP vs. number of growth rings per 0.5mm for two species of *Belemnitella.*

	Belemnitella langei					*Belemnitella minor* I						
DVDP (mm)	4.8	7.6	7.9	13.2*	M	10.1	10.2	11.1	16*	20	M	OM
P	24	25	24	-	24.3	22	21	19	-	17	19.7	22
C	19	18	17	-	18	16	16	15	-	14	15.2	16.6
O	16	15	14	-	15	13	14	12	-	11	12.5	13.7
TR (est)	94	130	164	210-230	-	179	173	169	230-250	280	-	-
TR (actual)	-	-	-	-	-	-	177	-	-	-	-	-

P = number of rings visible immediately adjacent to the protoconch per 0.5mm

C = number of rings visible in centre of the guard per 0.5mm

O = number of rings visible from outer edge of guard per 0.5mm

M = Median; OM = Overall Median (both species)

TR = Total number of rings (estimate of average growth rates)

* Estimate of total number of growth rings for a maximum-sized individual of the normal population. Note: a complete count was only possible on one specimen of *B. minor* I (DVDP 10.2mm). The difference between estimated and actual TR suggests an error of ± 2-3% in the results overall.

B. minor I appear to be present in any of the samples.

Table 1 shows that *B. langei* had a slower growth rate than *B. minor* I and attained a smaller overall size. Maximum recorded DVDP in this study was 12.3mm (although Christensen (1995) records one specimen with a DVDP of 13.2mm. Scaling up from the collected data it is possible to estimate that these individuals would have had in the region of 210-230 growth rings. Similarly the main population of *B. minor* I attains a maximum DVDP of 16mm (see Fig. 3a-d). The estimate for an individual of this size is a total of around 230-250 growth rings. If this were translated into days (see discussion above), the life spans of species of *Belemnitella* appear to be broadly comparable to

modern warm-water cephalopods. Schonfeld and Burnett (1991) estimated surface water temperatures of 17°C in the Upper Campanian *B. minor* zone of Norfolk which would place it in a sub-tropical climatic zone.

The "giant" specimens of *B. minor* I (DVDP 18-20mm) appear to form a separate population group and are only found in this section above Flint Band 10. A count of 17 rings per 0.5mm near the protoconch on a specimen (DVDP 20mm) demonstrates the fastest initial growth rate recorded in this study. As can be seen from Figure 3 these large specimens group outside the normal distribution curve of the main population. The estimate of 280 growth rings (when translated into days) also gives the greatest age detected. As mentioned above it is possible to speculate that these "giants" migrated into the area from slightly cooler waters.

This study should be regarded as a preliminary investigation. Clearly more work is required in analysing more material of different genera and different ages to obtain an overview of belemnite growth patterns. More accurate results (total ring counts) might be obtained from thin sections or etched specimens as described by Stevens and Clayton (1971).

Population dynamics

The relative age structure of the *Belemnitella* populations was analysed in two different facies (hardground and soft chalks). The exercise was carried out at a group level (specific identification of all specimens being impracticable) on 63 specimens from the hardground below Flint Band 7; 61 specimens from the *Echinocorys* Bed where 2 species are present and 97 specimens from between Flint Bands 10 and 11; and 20 specimens from the hardground above Flint Band 11 where 5 species are present.

Figure 3(a-d) shows the dorsal-ventral diameter at the protoconch (DVDP) in mm plotted against the number of specimens of each group. This is used as a measure of relative age (see discussion above). Areas where the DVDPs overlap have been superimposed on one another, so the size distribution can be read clearly. From this it can be seen that the populations vary significantly between the four horizons. It is also clear that the concentration of guards was not caused by a mass mortality spawning event, as immature individuals are represented in all plots and the assemblages are not monospecific. The possibility of sexual dimorphism having some effect on the size

distribution of the population should not be discounted, although this is unquantifiable at present.

In the hardground below flint band 7 (Fig. 3a) the *B. langei* group has a skewed distribution with few large specimens represented. The *B. mucronata* group is skewed the other way with only adult representatives. No truly small belemnites (DVDP 2-4mm) are found in this horizon. This is probably due to prolonged exposure, winnowing and boring on the sea floor.

The plot for the *Echinocorys* Bed (Fig. 3b) has a bi-modal distribution for the *B. langei* group. This is caused by the relatively large numbers of juvenile forms present. The *B. mucronata* group consists entirely of adult guards.

Between Flint Bands 10 and 11 (Fig. 3c) the *B. langei* group shows a nearly normal distribution, displaying a balanced population. The three peaks seen in the *B. mucronata* group distribution are caused partly by the presence of "giant" forms *of B. minor* I. Again the population is entirely adult.

The hardground above Flint Band 11 (Figure 3d) has a small total population. The distribution of groups is basically similar to that seen in the hardground below Flint Band 7.

The basic pattern seen in this analysis is that on hardgrounds the *B. langei* group and the *B. mucronata* group are mainly represented by mature individuals. This may be due to the loss of smaller specimens due to the action of boring organisms or dissolution and exfoliation. However, small representatives of the *B. mucronata* group are only rarely encountered throughout the section. They also tend to belong to the smaller species *B. pauli* (known range of DVDP 7.4-14.1mm) and *B.* sp. 1 (known range of DVDP 9.3-13.6mm) (Christensen, 1995) where specific identifications can be made. This age distribution is the norm for belemnite populations in off-shore chalks. Balanced populations containing both adults and juveniles are normally only found in near-shore environments (Christensen, 1976).

In the soft-white chalks the *B. langei* group has a normal distribution (except for the abundance of juveniles in the *Echinocorys* Bed, which may represent nearby spawning events) which suggests the Beeston Chalk at Caistor represented their normal habitat. The *B. mucronata* group is only represented by adult individuals. A possible explanation is that the latter group of species migrated to the Caistor area to feed, spawning taking

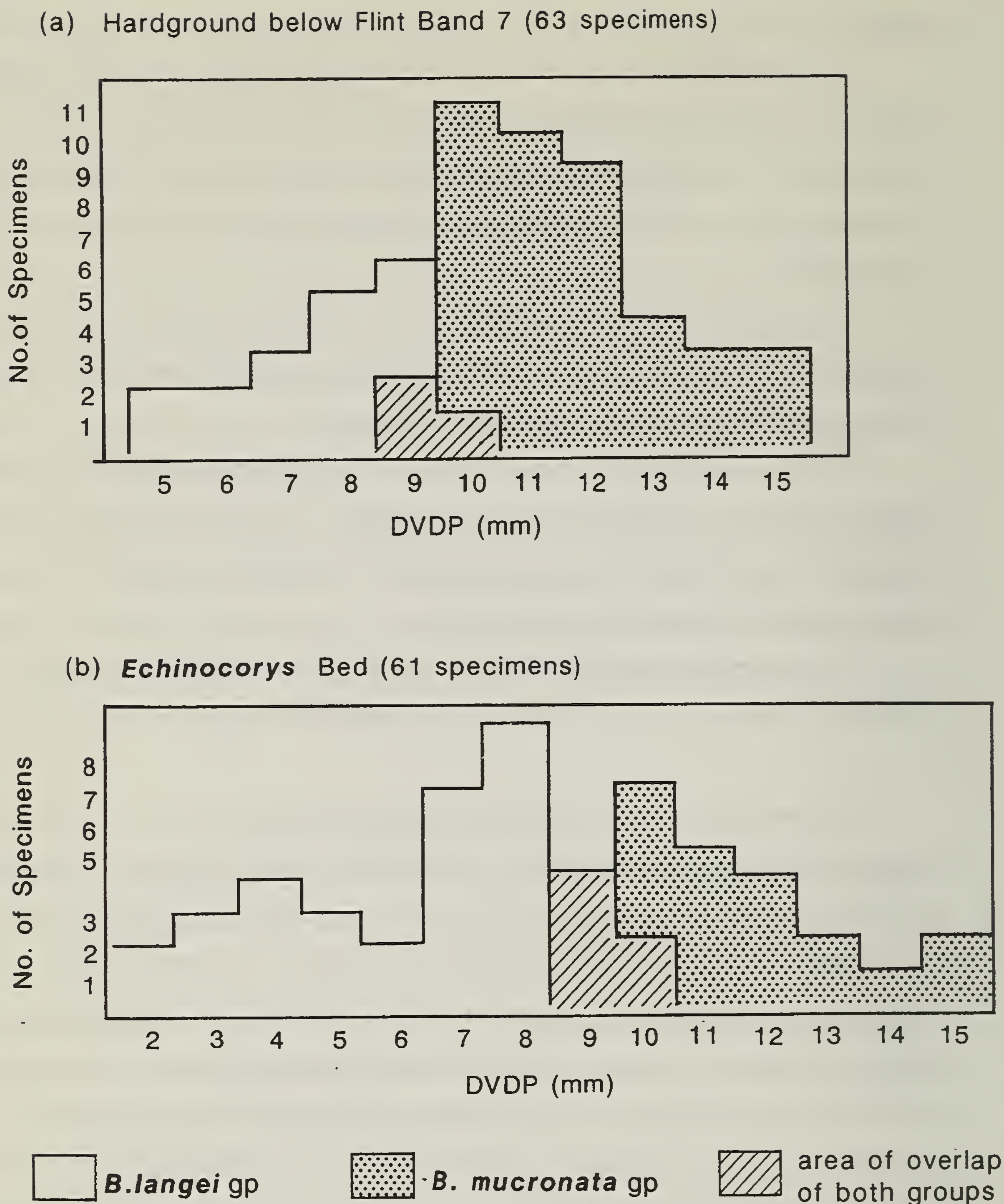
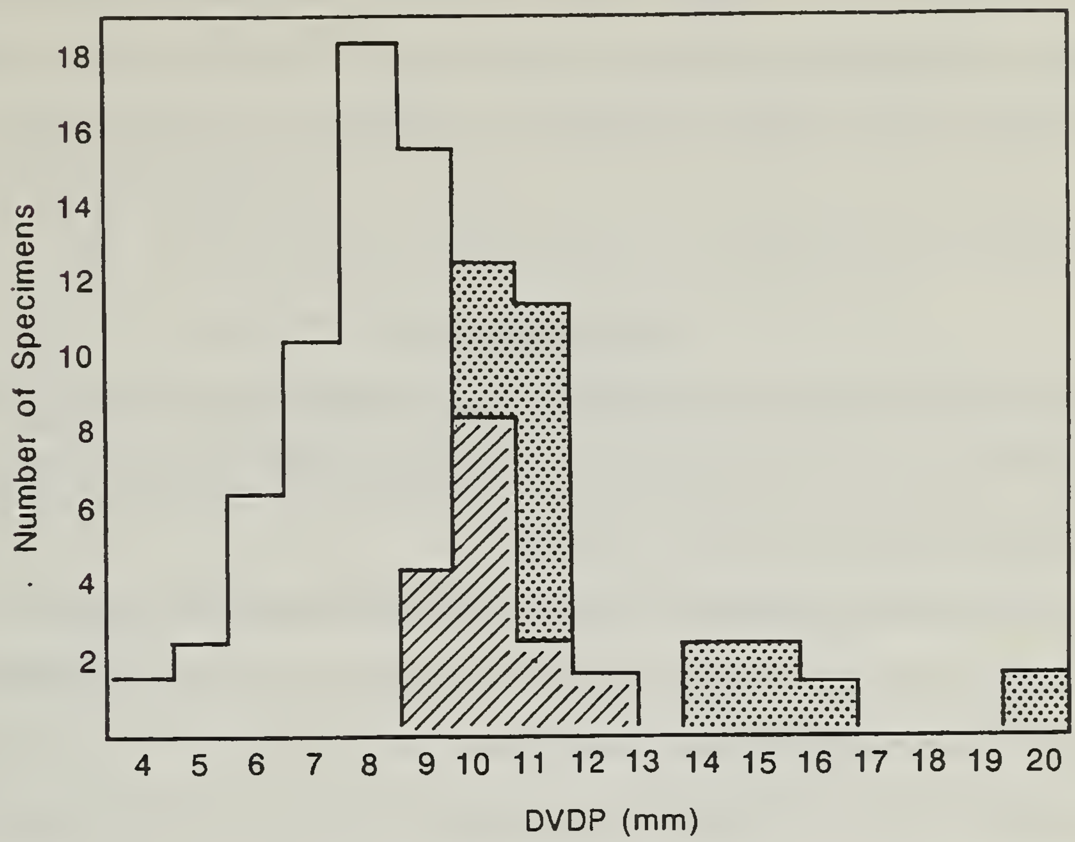
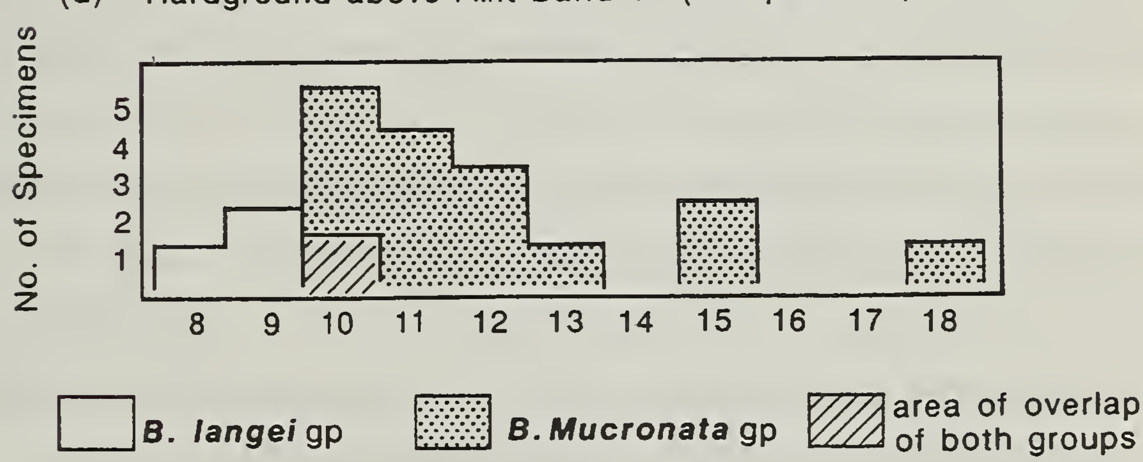


Fig. 3. Dorso-ventral diameter at protoconch versus number of specimens as measure of age distribution of the populations of *Belemnitella* in the Beeston Chalk: (a) Hardground below Flint Band 7; (b) *Echinocorys* Bed; (c) between Flint Bands 10 and 11; and (d) Hardground above Flint Band 11.

(c) Between Flint Bands 10 and 11 (97 specimens)



(d) Hardground above Flint Band 11 (20 specimens)



place elsewhere, possibly in near-shore environments close to the various small islands which dotted the shelf seas at this time (Christensen, 1976). The *B. langei* group, which appears to be endemic to the Caistor region, may have occupied a slightly different biogeographical/ecological niche as it appears to have bred in the general vicinity.

The switch in dominance between the *B. langei* and *B. mucronata* groups from soft chalk to hardground seems to display some cyclicity possibly related to sea level change. This, and the occurrence of juveniles in an offshore deposit, is unusual and requires further study.

ACKNOWLEDGEMENTS

I thank the students of City College, Norwich, Roger Cutting and Gawain Godwin who helped with the collection of specimens at various times. Thanks also to Paul Whittlesea and Peter Murphy who critically appraised the manuscript at an early stage and Julian Andrews, Andrew Pitchford and Peter Doyle for help with rewriting the earlier draft.

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A NOTE ON THE POSSIBLE OCCURRENCE OF *BELEMNITELLA* CF. *AMERICANA*? (MORTON) IN THE BEESTON CHALK, UPPER CAMPANIAN, NORFOLK.

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INTRODUCTION

Wood (1988) noted finding a "single example of an extremely elongate fusiform belemnite with a distinctly triangular cross-section, which may belong to the North American group of *Belemnitella americana* (Morton)" in the *Echinocorys* Bed of the Beeston Chalk Member (Upper Campanian) at Caistor St. Edmund. A single specimen was also found by Godwin (1998) in the same bed. At least two further examples were collected at the same locality by University of East Anglia undergraduates in 1990. These specimens were larger than the example figured here.

Attempts to identify the students' specimens to specific level at the time were unsuccessful - the overall morphology was suggestive of *Actinocamax* or even *Neohibolites* but the presence of a small mucron on one of the specimens suggested it was a species of *Belemnitella*. (*B. minor* and *B. langei* are the normal constituents of the belemnite population at this horizon). The specimen probably represents an early growth stage but is dissimilar to any other juvenile *Belemnitella* collected by Godwin (1998) in this section.

DESCRIPTION

The guard (Fig. 1) was collected, from just below the *Echinocorys* Bed, as five separate fragments; these could not be split to determine internal measurements. The guard is very slender, elongate and is fusiform (hastate) in shape. It flares slightly at the alveolus which was shallow and sub-conical in shape. No ventral fissure was noted on this specimen

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(which was slightly damaged at the alveolar end) but this was seen on another specimen. In apical view it is sub-cylindrical and in adoral view it is sub-triangular. The surface is smooth with no discernible ornament. The apical end features a small mucron. Its dimensions are as follows:

Length: 52.2mm

Maximum diameter: 4.5mm (near apical end)

Minimum diameter: 1.9mm (towards the alveolus)

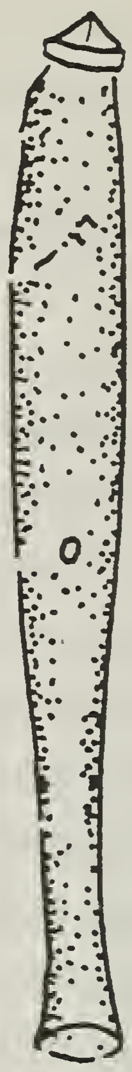

Fig. 1. *Belemnitella* cf. *americana*? (2.25 x natural size).
Note coral encrusted on posterior end.

The specimen is unregistered and held in the author's collection. It has been tentatively assigned to *Belemnitella* cf. *americana*? (Morton) on grounds of its overall morphology.

DISCUSSION

North American belemnites of the Late Cretaceous were reviewed by Christensen (1993). *B. americana* has a geographically and stratigraphically isolated occurrence in the Mount Laurel Formation and lower part of the Navesink Formation (Upper Campanian) on the Atlantic coastal plain near New Jersey where it is abundant (Owens and Sohl, 1973). Populations there consist of all growth stages and their normal palaeoenvironment appeared to have been a shallow, sandy inner shelf area with water depths of about 30m (Christensen, 1993; Owens and Sohl, 1973). The species' antecedents are uncertain. It

could be an endemic North American species but it is currently thought to be an ancestor of the *Belemnitella* lineage from the European platform (Christensen, 1993).

Specimens of *B.* cf. *americana*? were collected from an horizon just below the main concentration of echinoids in the *Echinocorys* Bed. They form less than 1% of the fossil population of belemnites in the bed. Whether their inclusion in the sediment took a matter of days or many thousands of years is unclear. If it is assumed that these belemnites were derived from the American population (which is far from certain) two possible hypotheses for their inclusion in the Beeston Chalk can be constructed.

1) A severe storm on the North American inner shelf caused a mass kill of the local population. Carcasses floated to the surface (when the phragmacone was unpunctured) and were carried by wind generated surface currents onto the European shelf where their remains decayed or were scavenged. The guards were then deposited (see Barthel *et al.*, 1990). This entire process might have taken only a few days (the Atlantic Ocean being considerably narrower in the Upper Cretaceous - by c. 1000 kms - than it is today).

2) A small shoal of adult belemnites was driven off the North American shelf by severe weather or some other factor. They reached the Caistor area on the European epi-continental shelf which provided a rich food supply for nektonic carnivores (Godwin, 1998). The adults spawned but their descendants were unsuccessful due to competition from the indigenous population or possibly unfavourable oceanographic conditions. In a short span of geological time this small migrant population became extinct.

Scenario 2 can be contrasted with the *B. langei* lineage which appeared in the Beeston Chalk near its base and continued to thrive until the end of the Upper Campanian. *B. langei* is thought to be a migrant from Eastern Europe (Christensen, 1995). It is clear that most long distance migrations (either pelagic or between basins) involving neritic species will be involuntary and accidental, although short distance migration between spawning and feeding grounds is generally assumed. Species which cannot successfully compete or adapt at a new location will be unlikely to leave any record of their presence in the rocks. Although the migration in scenario 2 ultimately failed, it demonstrates how belemnite species might appear at a site with no apparent evolutionary or geographic links. Such isolate populations can also rapidly evolve to produce new endemic species.

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PALAEOENVIRONMENTS AND CYCLICITY OF THE BEESTON CHALK (UPPER CAMPANIAN), NORFOLK AND THEIR POSSIBLE LINKS WITH THE NEKTONIC PALAEOECOLOGY OF *BELEMNITELLA*

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ABSTRACT

The palaeoenvironments of the Beeston Chalk at Caistor St. Edmund are examined in terms of their palaeogeography, palaeoclimatology and palaeoceanography. The area was probably influenced by nutrient-rich up-welling currents, which led to an abundance of pelagic life and a sparse benthos. The faunas and environments at Caistor in the Upper Campanian are comparable to those found on modern sea-mounts and off-shore banks. The evolution, palaeobiogeography and palaeoecology of ***Belemnitella*** *is reviewed and the facies distribution of the* ***Belemnitella mucronata*** *and* ***Belemnitella langei*** *groups is discussed. The concentrations of large numbers of individuals was probably caused by a combination of winnowing and predation.*

The changing faunas and facies within the Beeston Chalk also allows the demonstration of cyclic climate change. Meso-cycles in the order of 41-100,000 years and macro-cycles of 300-360,000 years are believed to have been detected. It can be shown that the dominant group of ***Belemnitella*** *at any given horizon is dependent on whether a cycle is in a transgressive or regressive phase.*

INTRODUCTION

Belemnite guards are very abundant in certain horizons in the Beeston Chalk. The possible causes of these fossil concentrations are the subject of this paper and are discussed in terms of belemnite population dynamics and predation. Water depth and

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temperature along with an overview of the probable ecosystems existing in the Chalk Sea at the time are also considered.

This study concentrates on a collection of fossil material (including over 300 belemnite guards) made in the Beeston Chalk Member (Upper Campanian Chalk of Norfolk) in the quarry at Caistor St. Edmund (Grid Reference TG 2390 0466). The Beeston Chalk occupies the *B. minor* I zone (Christensen, 1995) but a detailed study was only carried out in a 4.6m section of phosphatic chalk within sub-zones 2 and 3. Formerly this rock unit was placed in the lower part of the *B. langei* zone (Peake and Hancock, 1970). The lithostratigraphy and biostratigraphy of the Campanian chalks of Norfolk have been fully discussed by Jeletzky (1951), Peake and Hancock (1970), Wood (1988), Cox *et al.*, (1989), Johansen and Surlyk (1990), Pitchford (1991a; 1991b), and Christensen (1995), and are not discussed further here.

Detailed benthic palaeoecology was examined by Pitchford (1989) and Whittlesea (1996a; 1996b). The population structure of *Belemnitella* within the Beeston Chalk is discussed in Godwin (1998a).

The sequence of beds at Caistor St. Edmund is given in Figure 1. The strata selected for detailed study range from just below Flint Band 7 to Flint Band 12 of Wood (1988; see Fig. 1). This includes the *Echinocorys* Bed. Flint Band 7 is equivalent to Flint Band Z of Peake and Hancock (1970) and Flint Band 3 of Pitchford (1991a). The lowest half of the exposure, from Flint Band 9 downward, is now filled in or no longer accessible without risk.

PALAEOENVIRONMENTAL BACKGROUND TO THE NORTHERN MARGIN OF THE LONDON-BRABANT MASSIF

The Beeston Chalk was deposited in a warm, shallow, epi-continental sea around 74-73ma (Hancock, 1987). At that time the Upper Cretaceous transgression was reaching its maximum (Hancock and Kauffman, 1976), although this would have been accompanied by many minor oscillations in relative sea-level at a given site. Climatic conditions were vastly different to the present day, with a low latitudinal temperature gradient, which left the poles ice-free in what has been termed a "green-house" world (Doyle *et al.*, 1994).

The site at Caistor appears to have been positioned at the junction of two neritic (shelf sea) environments: the South North Sea Basin and the submerged London-Brabant Massif. This transitional position makes Caistor of great interest for palaeoecological

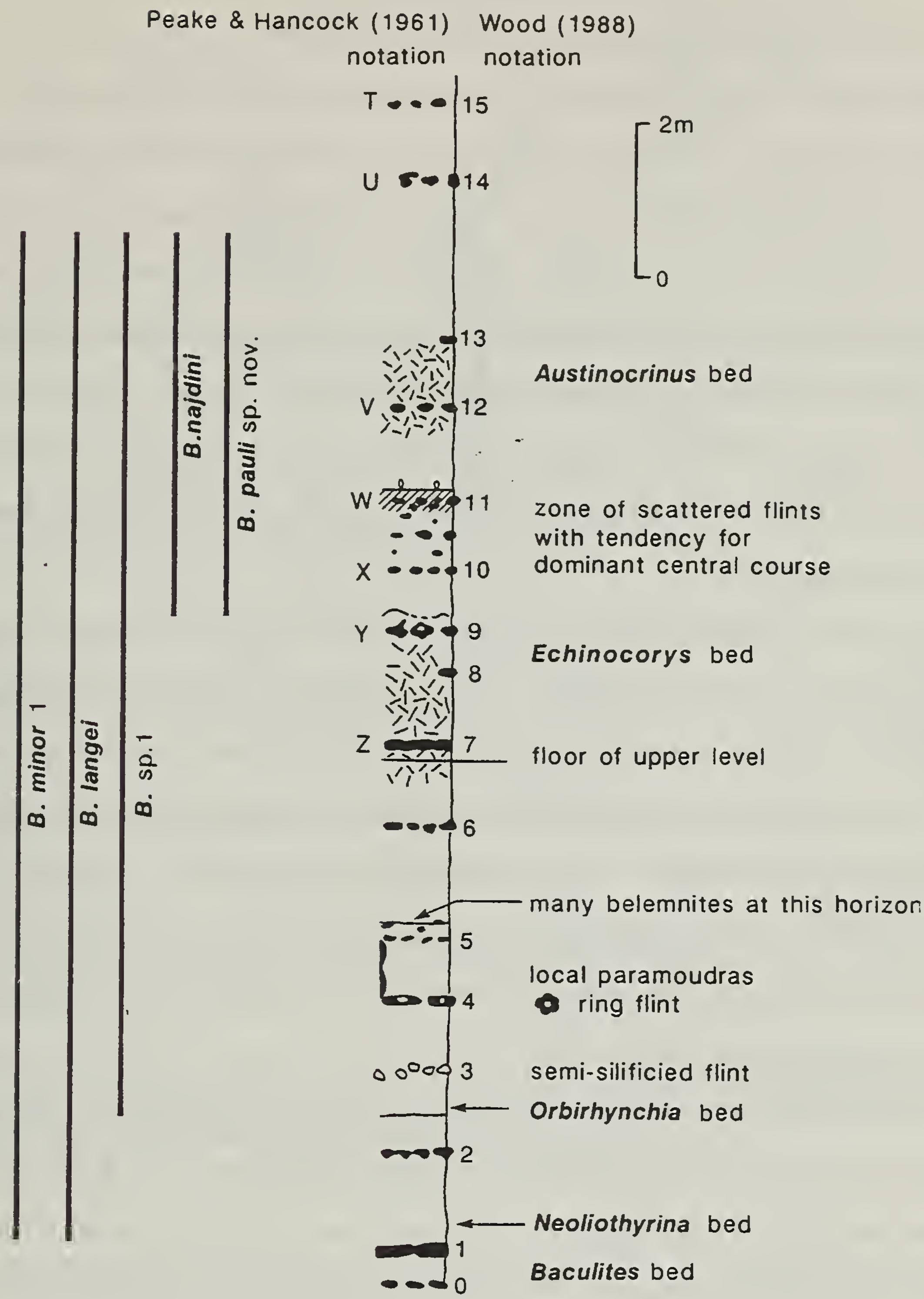

Fig. 1. Stratigraphy of the Beeston Chalk at Caistor St. Edmund and the ranges of *Belemnitella* spp. (Wood, 1988; Christensen, 1995). The hatch patterns (just below flint band 7 to flint band 13) represent chalk with abundant *Inoceramus* shell fragments (*Inoceramus* 'floods' of Wood (1988)).

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studies. The mosaic of basins, troughs, and submerged massifs which formed the environments of the Chalk Sea were probably responsible for the diversification and provincialisation of faunas noted below (Jablonski and Bottjer, 1983; Håkasson *et al.*, 1974).

Palaeogeography

A regional palaeogeographic reconstruction is given in Figure 2. This is based on maps and information derived from Hancock (1975), Cameron *et al.*, (1992), Doyle *et al.*, (1994), Tyson and Funnell (1987) and Smith *et al.*, (1994). Structural contour maps, isopachyte maps and borehole data from Cox *et al.*, (1989), Cameron *et al.*, (1992) and Arthurton *et al.*, (1994) show that Caistor, in Upper Campanian times, stood near the crest of a steep palaeo-slope on the northern margin of the London-Brabant Massif. This had a north-west to south-east axial trend, facing the deeper waters of the North Sea Basin. Upper Cretaceous sea-floor topography is unknown but the area may have simply been placed on the flanks of the massif or taken the form of a local high, forming a seamount or off-shore bank.

Sea-level was probably near its maximum height during the Upper Campanian (Hancock and Kauffman, 1976; Hancock, 1987). A shallow, warm epi-continental sea covered much of the western Eurasian continent from Ireland to the Russian Platform. An archipelago of eroded islands and the Fenno-Scandinavian Highlands provided very little sedimentary input into the region. This, and a warm, arid climate in western Europe, allowed the deposition of a remarkably pure marine chalk (Hancock, 1975). This eperic sea was bounded to the east by the Atlantic Ocean and to the south by the Tethys circum-equatorial Ocean (Tyson and Funnell, 1987).

Pitchford (1991a) estimated that some 125m of Upper Campanian Chalk has been preserved in the Norwich area compared to the greater thicknesses to be found in basinal and trough environments of the Southern North Sea, where the entire Upper Cretaceous may comprise over 1000m of sediment. This reflects a degree of condensation, but not as extreme as in near-shore sequences.

Palaeoclimate

It is thought that in the Upper Cretaceous global climates were warm with greatly expanded tropical and sub-tropical to warm temperature zones. *Belemnitella* is thought to have been temperature limited and is only rarely encountered on the northern margin of the Tethys Ocean (Christensen, 1976; Christensen *et al.*, 1990, 1993; Doyle, 1992) which some workers believe bordered a "super-Tethyan" climate zone (Kauffman and Johnson, 1988).

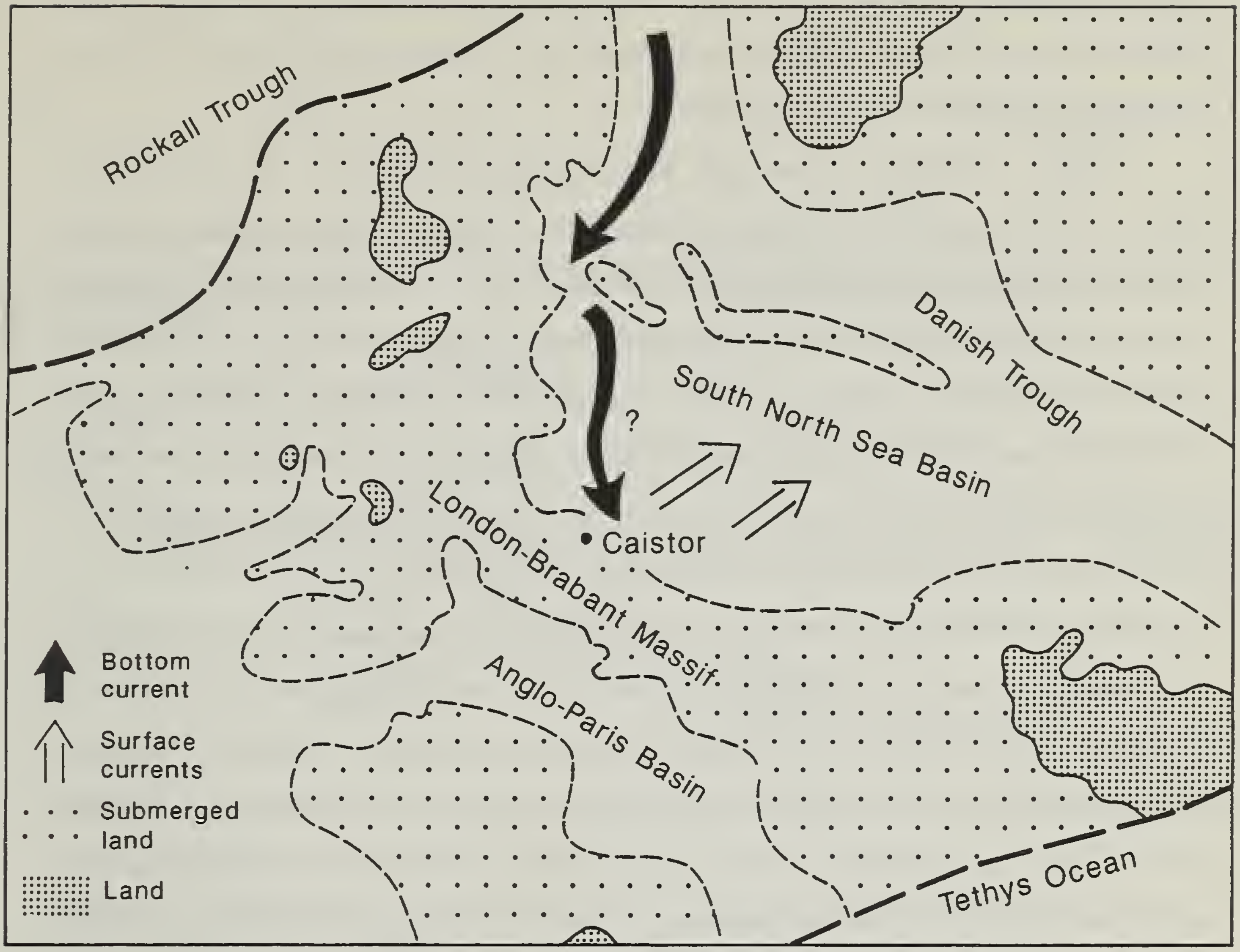

Fig. 2. Palaeogeography of the northwestern European shelf during Late Campanian times (after Tyson and Funnell, 1987; Hancock, 1975).

The small land masses of northwestern Europe probably had arid, sub-tropical climates but there is some evidence for seasonal rainfall from floral provinces and climate reconstructions (Batten, 1984; Robinson, 1995).

Palaeoceanography

Evidence for surface and bottom water sea palaeotemperatures comes from oxygen isotope geochemistry. Urey *et al.*, (1951) suggested 14.4 to 15°C for surface waters in the Campanian of Norfolk, based on the analysis of belemnite guards. However, Bowen (1961) thought surface temperatures to have been around 20°C in a study also based on belemnite guards. Temporal variations and the migratory habits of the individuals tested

could explain these variations. The work of Schonfeld and Burnett (1991) implied surface temperatures of 17°C at Laagerdorf (Germany) and Norfolk (surface water in the *B. minor* zone (uppermost Upper Campanian).

Climate modelling suggests that the prevailing wind direction over the London-Brabant Massif was NNW. Resulting wind-driven surface currents would have been directed to ENE due to the coriolis effect (Barron, 1985). The experiments of Luyendyk *et al.*, (1972) suggested that bottom currents may have been directed into the North Sea Basin from the North (see Fig. 2). It has been postulated that a proto-Gulf Stream had developed in the Upper Cretaceous and these currents could have been part of the return flow (Hart, 1976).

Up-welling and dysaerobia. Bottom currents are not likely to have been very strong and were probably intermittent. Oceanic circulation is believed to have been density driven in the Upper Cretaceous and in consequence bottom currents were sluggish (Research on Cretaceous Climates Group, 1986). As the winds would have been directing the surface current off the London-Brabant Massif, this could have led to an up-welling of nutrient-rich bottom water in the Norfolk area. Jarvis (1980a; 1980b) cites this as the major cause of phosphatic chalk deposition in the Upper Cretaceous. The sea-level maximum would also have led to higher organic productivity and an elevated nutrient supply. Under these conditions shelf sea-floors may become dysaerobic (Hart and Biggs, 1981), and provide conditions suitable for the development of phosphatic chalks (Jarvis, 1980b). The latter author linked this process to minor regressive episodes, which led to increased scouring by bottom currents and hardground formation in shallow water. In consequence the phosphates were winnowed and concentrated in the sediment. These episodes were followed by quiet, transgressive pulses where deposition occurred under dysaerobic conditions (see below). A reconstruction of proposed conditions during deposition of the Beeston Chalk is given in Figure 3.

Tyson and Pearson (1991) list criteria for the formation of ancient dysaerobic facies which are applicable to the Beeston Chalk. These include an abundance of amorphous organic-derived matter (phosphates in this case: a planktonic/nektonic fauna; benthos absent or of low diversity; low level of bottom current activity; warm palaeoclimate; and a cyclic pattern of deposition on a Milankovitch time-scale (see

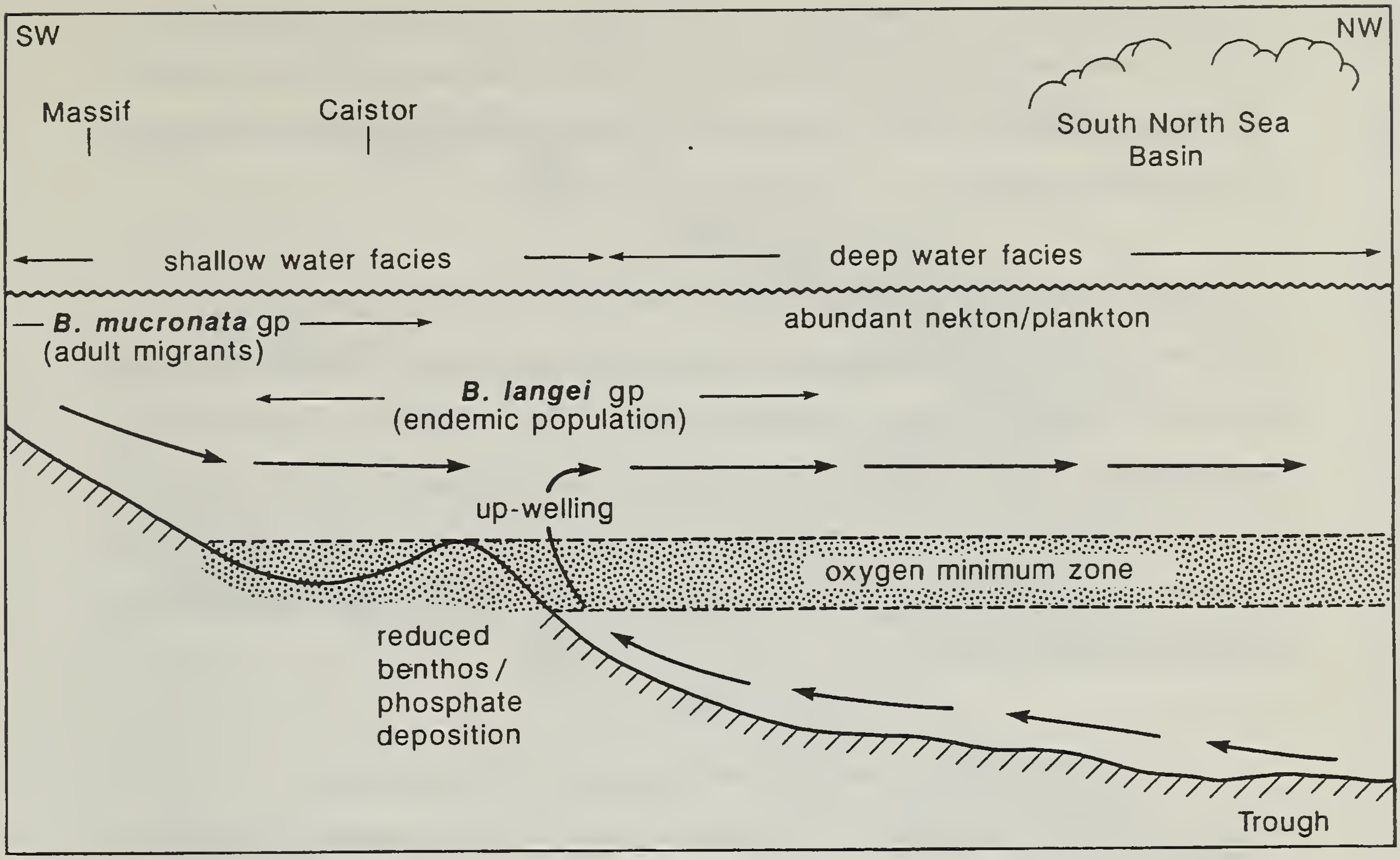

Fig. 3. Hypothetical cross-section of the palaeogeographic factors affecting the Caistor area during the Upper Campanian. (Not to scale)

below). They also state that such episodes are associated with minor regressive episodes during periods of high sea-level (see Cooper, 1977).

Bathymetry. Evidence from sponges and other fauna suggest the Chalk was deposited in water depths between 50 and 600m (Reid, 1968; Hancock, 1975), although a limit of 300m is thought likely for chalk now exposed on-shore. However, much of the chalk was deposited below the photic zone (Håkasson *et al.*, 1974). Hardgrounds such as those found in the Beeston Chalk, are initiated in sediments in water depths between 50 and 100m (Kennedy and Garrison, 1975).

Small, encrusting scleractinian corals occur in the Beeston Chalk up to Flint Band 9 suggesting deposition in the photic zone. This may exceptionally extend to more than 200m in depth but is frequently less (Doyle, 1996). Between Flint Bands 7 and 9 depths probably ranged from 100 to 150m based on the presence of small sponges and corals. Hardgrounds here are poorly developed, so shallower water depths, where development is

more complete, are not indicated (Kennedy and Garrison, 1975). From Flint Band 9 to 12 corals are absent, suggesting deposition below the photic zone, possibly in water depths of 150-200m. The related factors deduced from palaeogeography and palaeoceanography are summarised in Figure 2.

FAUNAL REALM AND FACIES DISTRIBUTION OF *BELEMNITELLA*

The faunal realms occupied by lineages of Belemnitella in the Upper Cretaceous have been discussed by Naidin (1959), Stevens (1963; 1973), Christensen (1976; 1991), Combémorel *et al.*, (1981) and Doyle (1992). The area occupied by the British Isles in the Late Cretaceous has been defined by Christensen (1976) as part of the Central European Sub-Province of the North Temperate Realm, which is characterised by the family Belemnitellidae PAVLOV.

The evolution and palaeobiogeography of *Belemnitella*

To help understand the speciation of *Belemnitella* in the Upper Campanian of Norfolk it is useful to review the evolution of this genus (as summarised in Fig. 4). It is clear that evolution proceeded mainly as a series of radiations from the Russian Platform. In the Upper Campanian this pattern of speciation broke down and became provincialised.

There is a repeated trend for large species to evolve through time with minimal morphological change (e.g. *B. mucronata* > *B. woodi* > *B. minor*) with minor off-shoots producing smaller, more slender species or sub-species (e.g. *B. minor* > *B. pauli*), whose similarities to the parent stock have created a taxonomical minefield. The genus, therefore, provides a fine example of phyletic gradualism or stasis, depending on viewpoint.

In Russia, numerous local species developed from *mucronata* stock (Jeletzky, 1955a). Similarly, new local species and sub-species (see Christensen, 1995) developed in the Weybourne and Beeston Chalks of Norfolk - *B. woodi*, *B. minor* I, *B.* sp. 1, *B. pauli*, *B. minor* II and *B. minor* III (all in the *B. mucronata* group). Of these *B. woodi* appears to have migrated to the Anglo-Paris Basin, where it occurs rarely, as does *B. minor* II. *B. minor* I appears to be confined to Norfolk but a questionable occurrence is cited in Poland (Christensen, 1995). *B. minor* III appears to originate in Norfolk and radiates out to Western Europe. Its appearance in the Baltic region is doubtful and occurrences there are probably local species allied to the *B. langei* group.

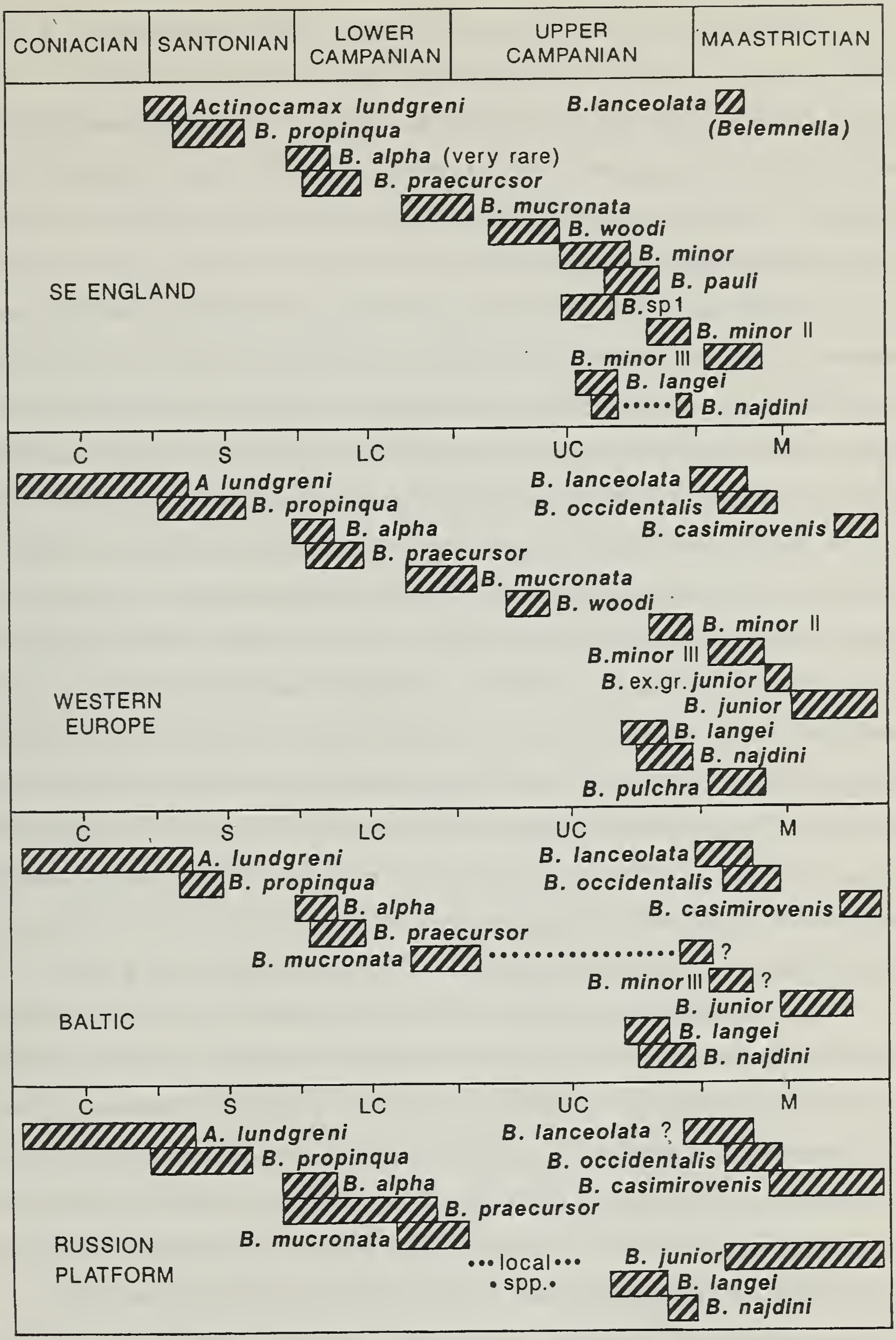

Fig. 4. The evolution of *Belemnitella* in the North European Province. Compiled from data in: Jeletzky (1955), Birkelund (1957), Christensen (1971; 1975; 1979; 1991; 1995), Christensen *et al.*, (1975), Christensen and Schmid (1987), Schulz (1982), Keutgen and Van der Tuuk (1990).

Christensen (1976) also offered an explanation for the provincialisation of faunas during the period. Although the discussion is about genera in the Late Coniacian - Early Santonian, the argument is equally applicable to differentiation within *Belemnitella* during the Late Campanian transgression. From this and Vbra's (1985) evolutionary theories and speciation hypothesis, as well as the palaeogeography given by Naidin (1959), the following hypothesis can be offered.

The Russian Platform offered a large basinal area with an abundance of varied, near-shore conditions which were ideal habitats for belemnites. The large basin offered fairly stable conditions. To the west the Cretaceous Chalk Sea was broken up into a mosaic of small basins and massifs. During transgressive pulses these would have become unstable (Doyle, 1992), leading to dysaerobic bottom conditions and disruption to the food web as discussed above. This could lead to local extinction by habitat tracking (i.e. changing environmental conditions move the preferred habitat of species to a new location forcing the population to move with it) or true extinction (if they are unable to migrate).

Environmental change, such as transgressive sea-level rise, causes habitat disruption. The up-welling proposed at Caistor, for instance, may have caused the benthic communities to be largely excluded by dysaerobia, disrupting the food webs found at the base of the Beeston Chalk (see below). However, the elevated nutrient supply caused increased primary productivity in the plankton which in turn attracted nektonic plankton feeders and their predators, creating new habitat possibilities. Surviving belemnite species may have fragmented to take advantage of the ecological niches that were created.

After this radiation event the fauna developed locally with only a small overlap with other belemnite populations on the north-west European shelf. Competition for the same ecological requirements would have impeded intermingling at a later stage. There is good evidence that populations were separated by the deeper basins as few belemnites are found in basinal chalks (Christensen, 1976). This is unlikely to be due to bathymetry. A more probable cause is lack of a dependable food source. Climatic and oceanographic zonation may have played some part as well. Essentially the appearance of many new species at Caistor, and elsewhere, can be seen as a direct consequence of the Mid-Campanian transgression.

Facies distribution

As belemnites are generally very uncommon in off-shore chalks, the number of specimens in the Beeston Chalk at Caistor St. Edmund can be viewed as unusual. As nektonic predators, belemnites are linked to the marine food web (Christensen, 1976). Surlyk and Birkelund (1977) regarded the deposition of guards in off-shore chalks as being merely accidental, as the individuals are far from their normal near-shore habitats, where populations are more numerous with assemblages of both young and mature specimens. One possible mechanism for this is the post-mortem drifting suggested by Barthel *et al.*, (1990). Another mechanism is occasional mortalities during migration. Most concentrations take place near feeding and breeding grounds which are generally in-shore. Typical facies include glauconitic sands and shallow marine carbonates and marls.

Caistor's intermediate position between the North Sea Basin and the London-Brabant Massif was probably broadly equivalent to a seamount environment providing an "island" of diversity in the shelf sea (Rogers, 1994). The deposition of phosphates, which were probably caused by up-welling, suggests a nutrient-rich habitat, and the evidence of relatively abundant pelagic fossils of nektonic predators in the vicinity (mosasaurs, sharks, fish and belemnites) (Wood, 1988) imply it was a good feeding ground. In contrast the benthos was relatively sparse.

The population dynamics of *Belemnitella* discussed in Godwin (1998a) show that only mature or near mature individuals of the *B. mucronata* group are present between Flint Bands 7 and 12 of the Beeston Chalk. This suggests that adults were migrating into the area from more near-shore habitats to feed and that their spawning grounds lay elsewhere. Two growth rate components to their population have been tentatively recognised which may relate to a reproductive strategy or the mixing of two populations, the larger size possibly coming from cooler waters which promotes "giantism" (Collins *et al.*, 1995; Forsythe and Van Heukelen, 1987).

The *B. langei* group is represented by juveniles and adults (Godwin, 1998a), which suggests that the Caistor area was their normal habitat especially during periods when the facies is indicative of deeper waters. It is conceivable, given the numbers of juveniles at some horizons (Godwin, 1998a), that they could have spawned in the vicinity.

Therefore, it is possible to speculate that the *B. langei* group represents an adaptation to a more epi-pelagic (sea-going) mode of life. As seen above it is widely

distributed (in a series of variant forms) across the Province at a time when the *B. mucronata* group is confined to a number of local populations in the various basins of the European continental shelf. The latter group probably preferred the more typical shallow water habitats offered by the massifs with their archipelagoes of emergent land, although both groups' biogeographic ranges overlap.

The fact that the *B. langei* group is rarer on omission surfaces and few of its juveniles are present (Godwin, 1998a) suggests habitat tracking during a regressive sea-level change; its spawning sites possibly moved off-shore. At other times belemnites are rare in both shallow and deep-water facies suggesting unfavourable environmental conditions forced populations to move elsewhere.

Jarvis (1980c) and Christensen (1976) suggested that the concentration of belemnites in the vicinity of omission surfaces may be causally linked through the food web. The firm substrate allowed for a greater and more diverse benthos, which in turn attracted nekto-benthonic predators. Nektonic predators in turn followed their prey. The increased productivity indicated may also be linked in the same fashion to the abundance of belemnites at Caistor. Most of the concentrations there appear to be predator regurgitates (see below), and the animals would have been consumed nearby. Strata containing rare, scattered individuals may just be the final deposition site of animals that had suffered post-mortem drifting. This could conceivably involve considerable distances (Barthel *et al.*, 1990).

Palaeobiology of belemnites

The biology of modern coleoids has been nearly as poorly understood as that of the extinct belemnites. In consequence ecological models were based on a poor understanding of coleoid physiology. The mode of life of belemnites was recently reviewed by Doyle and Howlett (1989). These authors conclude that belemnites were similar in form and function to neritic squid although some genera such as *Hibolites* appear to have adopted a more migratory oceanic life style.

Coleoids have been described as being genetically programmed to "live fast and die young" (O'Dor and Webber, 1986). All coleoids have a basically semelparlous reproductive strategy (spawn once and die) unlike the long-lived *Nautilus* which spawns repeatedly. A life span of 200-300 days has been proposed for *Belemnitella* by Godwin (1998a).

Like oceanic squid, *Belemnitella* appears to have had a javelin-shaped morphology adapted for manoeuvring in open waters (Boycott, 1965). On the other hand, their phragmacone would have limited them to shallow depths. The experiments of Westerman (1973) showed that belemnite phragmacones (for a range of species) imploded at 250-400m depth. These figures seem reasonable for Upper Cretaceous belemnites as the epi-continental shelf sea in which they lived probably varied between 100 and 300m in depth. The exceptions to this were the deeper troughs where depths could reach 600m (Jablonski and Bottjer, 1983; Håkansson *et al.*, 1974).

Belemnites were therefore apparently perfectly adapted to a predatory existence in shallow pelagic shelf seas. Another consequence of their overall morphology and retention of the phragmacone is that they were unlikely to have had a nekto-benthonic habit like the flat-bodied *Sepia* or the octopus.

PALAEOECOLOGY OF *BELEMNITELLA* IN THE BEESTON CHALK

The assemblage data and population dynamics for the belemnites described below are available in Godwin (1998a, figs. 2 and 3a-d).

Hardground below Flint Band 7

Most of the guards derived from the hardground below Flint Band 7 are dominated by the *B. mucronata* group (70%) with some "giant" specimens present (Godwin, 1998a). These guards show signs of prolonged exposure on the sea floor. The chalk here is indurated with many small nodules. This is the second incipient stage of hardground formation (Kennedy and Garrison, 1975).

The belemnite guards are a pale buff colour and may display exfoliation of the layers of the guard; multiple sponge borings, and encrustations by bryozoans and small scleractinian corals. These features range in intensity from complete coverage of the upper surface to very minor damage, indicating variable periods of exposure (see Figure 5a). The taphonomy of belemnite guards was studied in detail by Pitchford (1989).

Doyle and MacDonald (1993) suggested that corrosion from a predator's gastric fluids could affect the guards, leading to possible exfoliation. However, these features (as well as borings and encrustations) are always concentrated on one side of the guard making this unlikely. This also suggests that the rostra were unmoved for a considerable period after coming to rest, as disturbance would have led to the guards being bored on all

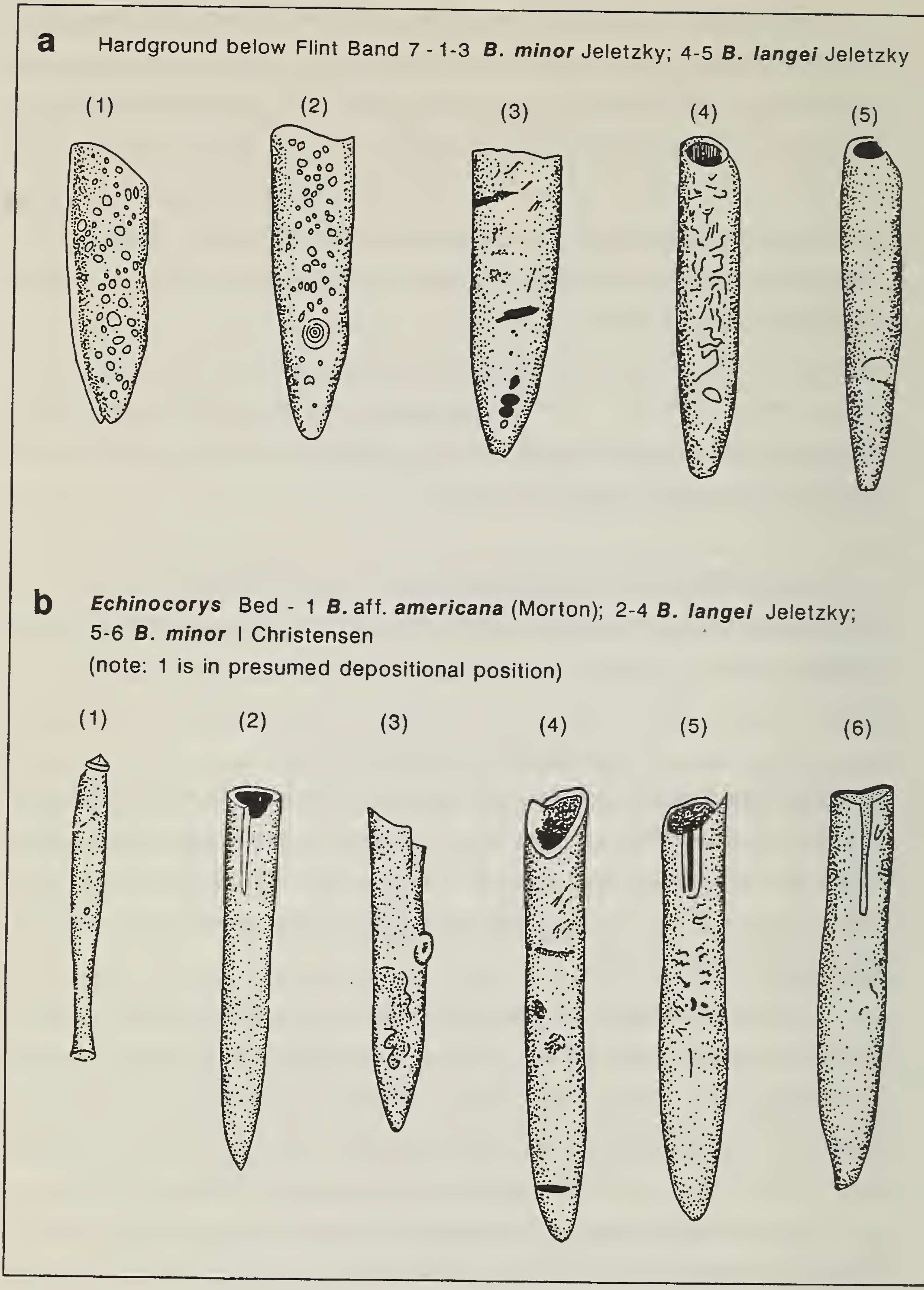

Fig. 5. Belemnites from the Beeston Chalk, Upper Campanian (note predator bite marks, borings and encrustations). Sketches are reproduced at 2/3 actual size.

sides. Most of the guards are fragmented and borings often occur on broken faces. The concentration of damage to one side suggests bottom currents were not strong enough to move the guards and demonstrates that breakage had occurred prior to deposition.

Microscopic examination of over 200 guards (including 41 from this hardground) revealed that many rostra had single or parallel striations on them, up to 40mm apart. These striations are at high angles relative to the long axis of the guard. One end of these marks characteristically has a small indentation which leads into the groove. This shallows, and tapers away after a short distance. Experimentation with a sharp awl showed that similar marks could be made with an initial tap, followed by a steady forward pressure, with the guard held in a vice. A heavy enough blow would fracture the guard. This is compatible with the marks being caused by teeth and that vertebrate predation was the main cause of mechanical fracture.

A number of large predators are known to have hunted in the Chalk sea of the Upper Cretaceous. Shark teeth are a fairly common fossil in the Campanian Chalk of Norfolk: although none were recovered during this study they have been found by other workers (Pitchford, 1989). However, a tooth fragment from a small mosasaur was recovered from the hardground. Mosasaur teeth, pelvic bones and vertebrae have also been collected from the Beeston Chalk from St. James's Pit Norwich (Benton and Spencer, 1995). They regard the mosasaur as the top predator of the Cretaceous seas. Comparison of this tooth fragment with material at the Castle Museum in Norwich suggests the specimen to be *Leiodon anceps*, a term which encompasses a number of species (Milner, 1987). Pliosaurs, elasmosaurs and marine crocodiles are also known from Campanian sediments (Siverson, 1992) but are likely to have preferred a near-shore habitat. There are no records of them in the Norfolk region.

Lamniform and squaloid sharks recently described for the Campanian and Maastrichtian of Sweden by Siverson (1992;1993), are a possible candidate as belemnite predators as their remains also occur in the Campanian Chalk of Norfolk. Siverson (1992) has speculated that some shark species were specifically adapted to prey on belemnites. In the Lower Campanian of Sweden, lamniform shark teeth are invariably only found in association with guards of *Belemnellocamax m. mammillatus*. The distance between striations on the Norfolk specimens are of the right spacing to fit the dentition patterns of these sharks (20-40mm).

Lamniform sharks (e.g. the Tiger Shark) are fairly large predators. Modern species of squaloids are smaller deep-water animals. They inhabit the bottom of tropical to temperate continental slopes, feeding mainly on small fishes, squid and crustaceans (Compagno, 1984). Fish predator bite marks on belemnite guards are also noted by Gall (1983).

Other fossil remains found in the hardground sediments are endo-benthonic and mainly preserved as shell debris. A mature community is indicated by inoceramid fragments and echinoid plates (both regular and irregular) which are widely scattered throughout the sediment. A single ossicle of the crinoid *Austinocrinus bicoronatus* was also recovered. An almost complete (but crushed) valve of *Inoceramus balticus*? has most of its ventral margin missing (this would have protruded from the sediment when in life position); this suggests predation by a bottom dwelling predator such as *Ptychodus* (a blunt-toothed shark) described by Kauffman (1972). This might explain the abundance of shell debris in an otherwise deep-water, low-energy environment.

The inoceramid "floods" (Wood, 1988), where inoceramid fragments are extremely abundant, begin just below Flint Band 7 (Fig. 1) in the Beeston Chalk. These "floods" may be recording the decline of the inoceramid populations which led to their extinction in the Maastrichtian by over predation (see Ward, 1992).

In summary, a low sedimentation rate combined with a low-energy environment seems to have prevailed during this stratigraphic interval. The belemnites appear to have been accumulated over a considerable period of time, most probably as predator regurgitates, as many specimens have several generations of sponge borings pitting their upper surfaces (see Fig. 5a). Low sedimentation rates are indicated not only by multi-generational colonisation of the guards by the epi-fauna but also by the formation of the hardground itself (Kennedy and Garrison, 1975). Guards that are better preserved were probably deposited later, and typically have small, ring shaped, coralline encrustations on them. This style of encrustation is more typical of the chalks within the *Echinocorys* Bed and implies that deposition occurred within the photic zone (maximum depth around 200m, Doyle, 1996). The hardground benthic assemblage probably represents a mature community (Jablonski and Bottjer, 1983; Whittlesea 1996b).

Echinocorys Bed

The chalk here is softer, yellowish in colour when weathered, and flecked with phosphatic pellets of faecal origin (Jarvis, 1980a). The rather sparse microfauna (<125 μm), although not examined in detail, contains large benthic foraminifers, ostracods and rare micromorphic brachiopods. The macrofauna also lacks diversity, being dominated by abundant fragments of inoceramid shell. This fauna represents a succession of pioneer communities, which only have small encrusters, apart from the horizon containing the majority of the *Echinocorys* which is a mature soft-bottom community. Here *Acutostrea* sp. and other oysters were present in small numbers with rare brachiopods, mainly *Cretirhynchia aff. norvicensis* and occasional examples of the echinoids *Galerites vulgaris* and *Cardiaster granulosus*. The community is illustrated in Figure 6.

The large specimens of *Echinocorys aff. conoidea* are found along a discrete horizon that must relate to a slightly higher bedding plane due to their infaunal habit. Gall (1983) suggested these irregular echinoid dominated beds are indicative of deeper water conditions (also see Felder, 1986). No juveniles of this species were found here and it is possible that the environment favoured adults. "Nests" of young are found higher in the sequence in Flint Band 11 (Wood, 1988).

Belemnites are much better preserved in these beds, having a fresh appearance and a translucent red-brown colour (Figure 5b). Encrustations on the guards, if present, are confined to a single generation and include sponges, corals and bryozoans. Many fragments had been broken prior to deposition and have identical scratch marks to those seen in the hardground below Flint Band 7, indicating predation. The *B. langei* group was dominant in this bed (59%) but some large specimens of *B. minor* I are present as well (Godwin, 1998a).

Entire guards, especially of juveniles, are easier to collect in this bed. Coral encrusters on one juvenile suggests that the sediment was thixotropic (see Fig. 5b-1); this guard appears to have entered the ooze vertically, posterior end upwards as a small coral can be seen encrusting its tip. Later, the guard was presumably tipped over, possibly by the action of burrowing or ploughing invertebrates, as it was recovered in horizontal position. A similar instance is recorded by Barthel *et al.*, (1990), and they regarded this as an indicator of very low-energy conditions. These authors suggested that necroplanktonic transportation is important; after death the body floats in surface currents. Air trapped in

a **Belemnitella langei**
b **Belemnitella langei**
c **Echinocorys** aff. **conoidea**
d **Belemnitella** aff. **americana**
e Teleost fish
f Hexactinellid sponge
g Hexactinellid sponge
h **Inoceramus** sp.
i Inoceramid fragments
j **Echinocorys** aff. **conoidea**
k **Cretirhynchia norvicensis**
l **Galerites vulgaris**
m **Thalassinoides**
n **Acutostrea** sp.

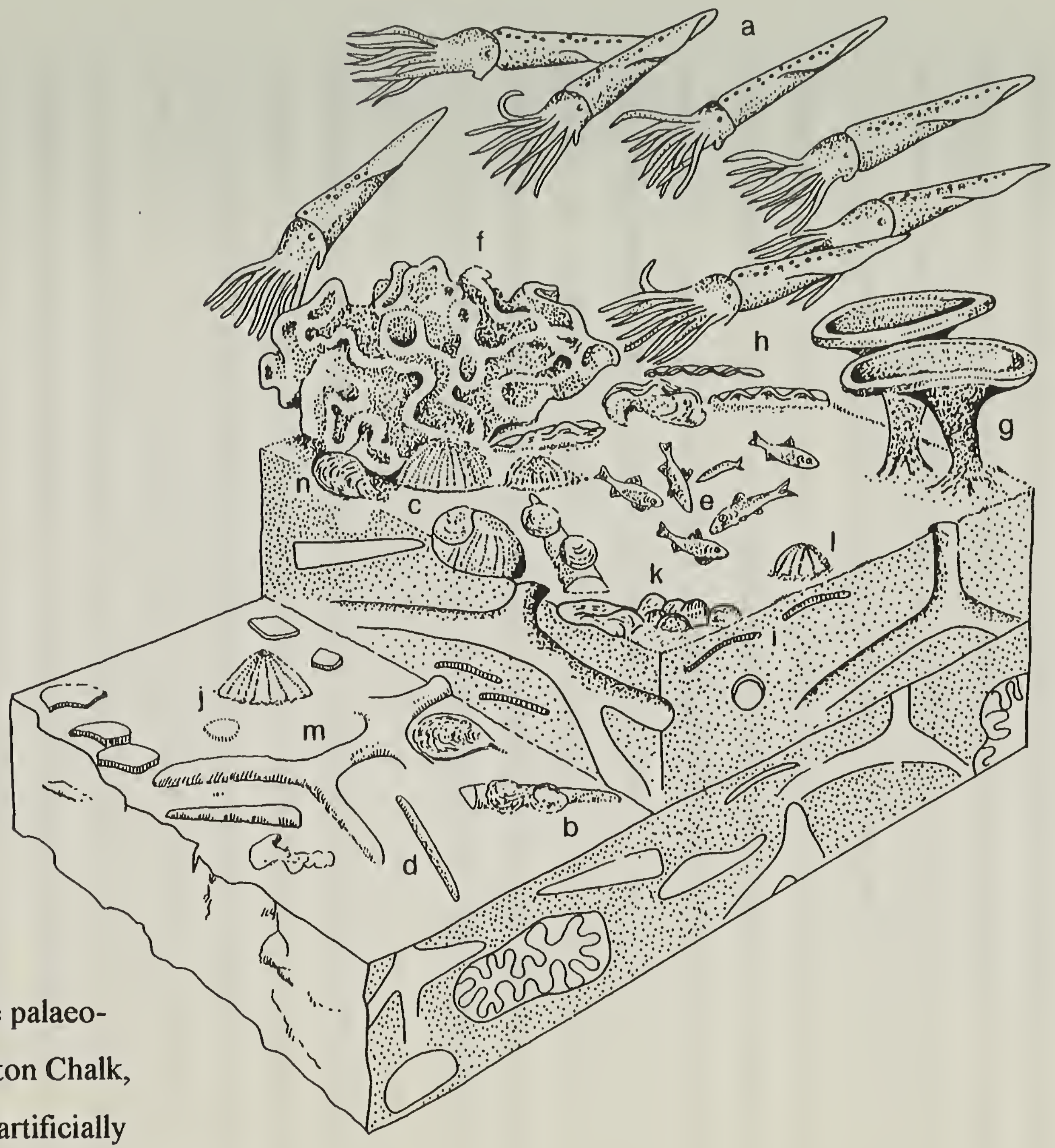

Fig. 6. Hypothetical reconstruction of the palaeo-community of the *Echinocorys* Bed, Beeston Chalk, Upper Campanian. Density of organisms artificially compressed. Redrawn and modified from McKerrow, (1978).

the phragmacone keeps it buoyant, until eventually the corpse disaggregates with the hard parts dropping through the water column to settle in the sediment.

The guard described above has a triangular cross-section and a smooth, very elongate, fusiform shape. It has been tentatively assigned to *B. aff. americana*? (Morton) see Godwin (1998b). Wood (1988) records a similar occurrence from the same bed. This belongs to a North American group and may have been carried into the Caistor area by a freak storm event (Godwin, 1998b).

The lower number of individuals per unit volume of rock in the *Echinocorys* Bed compared to the incipient hardground below Flint Band 7 (see Godwin, 1998a, fig. 2) is explained by a higher sedimentation rate with stratigraphic condensation occurring on the hardground. (The collection area of quarry face was c. 50 m^2 [*Echinocorys* Bed] versus c. 2.5 m^2 on the hardground). The higher sedimentation rate is demonstrated by the presence of a single generation of borings or encrustations on the *Echniocorys* Bed guards, as compared to the multi-generational borings seen in the hardground below Flint Band 7. The change in population structure shown in Godwin (1998a, fig. 3b), with the *B. langei* group becoming dominant (a cycle repeated further up the section), is probably a result of environmental change (as discussed below).

Incipient hardground above Flint Band 9

An incipient hardground is evident as a discontinuous, gently undulating, limonised surface. The exposure is intermittent forming part of a new trackway cut which is partially obscured by eroded chalk debris from higher levels in the quarry. Fossil material at this horizon is sparse and the diversity is low. Belemnite guards are rare, but show the same degree of prolonged exposure on the sea floor as the hardground below Flint Band 7. Exfoliation and multiple borings were common on the few guards recovered. The assemblage is dominated by the *B. mucronata* group (82%) (Godwin, 1998a, fig. 2).

Chalk between Flint Bands 9 and 10

The fauna remained very sparse in the phosphatic chalk at this level. Belemnites are rare but generally in good condition, with only a single generation of sponge borings or tiny coral encrustations. The assemblage is dominated by the *B. langei* group (70%) (Godwin, 1998a, fig. 2). The endobenthos included rare *Echinocorys aff. conoidea* and common fragments of *Inoceramus*. Epifaunal elements included occasional fragments of

Acutostrea sp. and other oysters and the brachiopod *Cretirhynchia norvicensis* is relatively common. The communities represented are probably successions of pioneer and mature faunas (as represented by the inoceramids - pioneers; and oysters and brachiopods - attached species of a mature community). However, environmental conditions do not seem to have been as favourable for the nekton as in the chalks below or above as the total abundance of the fauna is low. This may have been due to unstable oceanographic conditions leading to fluctuations in the nutrient supply and therefore oscillations between aerobic and dysaerobic bottom waters which would have disrupted the food web.

Chalk between Flint Bands 10 and 11

The chalk here is marked by an increase in abundance and diversity of belemnites. The *B. langei* group is dominant (65%) and the *B. mucronata* group included 10% of "giant" individuals (Godwin, 1998a). Species present include *B. langei*, *B. najdini*, *B. minor* I, *B. pauli* and *B.* sp. 1. Some guards were found in discrete, spherical clumps 30-50cms across; many guards were near entire, and striations which are thought to be predator bite marks were also observed (see above). This distribution of guards has been interpreted, by the author, as being caused by predator regurgitation.

Bivalves are poorly represented in this part of the chalk with inoceramid fragments and *Acutostrea* sp. being comparatively rare. The brachiopod population is however relatively abundant and diverse with *Cretirhynchia arcuata*, *Carneithyris carnea* and *Kingena* sp. present. These attached species represent a mature soft-bottom community. These faunas require a pioneer community (of inoceramids etc.) to have preceeded them. The pioneers would have provided shell debris to which the brachiopod spats could attach themselves, as their byssus will not have bonded with the coccolith ooze which was present on the sea-floor at the time. Most of the brachiopods in the chalk are micromorphic: the evolution of extremely small-size allows species to utilise tiny shell fragments (which the adult will not outgrow) as attachment points in a thixotropic environment (Johansen and Surlyk, 1990). Encrusting corals are totally absent here and do not reappear in the studied section. This indicates that deposition occurred below the photic zone (c. 150-200m +) as discussed above.

Hardground including Flint Band 11

The chalk here is indurated and an undulating omission surface is well defined. Belemnite rostra found at this horizon show typical signs of exposure for lengthy periods on the sea floor. Exfoliation, multiple sponge borings and encrustations are common. The assemblage is dominated by the *B. mucronata* group (80%), although the population is not as numerous as that found in the hardground below Flint Band 7 (Godwin, 1998a).

The endobenthos was developed at this level into a succession of mature soft to firm ground communities. Small specimens of *Echinocorys aff. conoidea* are found in Flint Band 11 where they are abundant. Inoceramid fragments are rare and the fragile remains of some hexactinellid sponges were also noted. The trace fossils *Thalassinoides* and *Zoophycos* were found preserved in several flints of Flint Band 11 itself, suggesting deepwater facies (Frey and Seilacher, 1980).

Chalk between Flint bands 11 and 12

Belemnites become relatively rare above the hardground of Flint Band 11 and once again the *B. langei* group dominates (57%) (Godwin, 1998a). Just above the hardground large specimens of *Echinocorys aff. conoidea* are found in a discrete horizon like the *Echinocorys* Bed but are not so numerous here. Above this bed fossil material is rare until just below Flint Band 12 where another inoceramid "flood" occurs. The overlying *Austinocrinus* Bed was not investigated as it was poorly exposed.

Palaeoecological causes of the belemnite accumulations

Doyle and MacDonald (1993) recently published a conceptual model to explain accumulations of belemnite rostra which they termed "belemnite battlefields". Such deposits are rare in the Upper Cretaceous chalks, mainly occurring on hard grounds as re-worked lag accumulations. Well known examples are found at Catton Grove, Norfolk, just above the Catton Sponge Bed (Wood, 1988); and at the base of the Vijlen Chalk Member in the Epen-Beutenaken area in the Maastrichtian of Belgium (Keutgen and Van der Tuuk, 1990). The concentrations discussed above do not deserve such an appellation. However, the model is equally applicable in explaining their formation.

Doyle and MacDonald (1993) listed five main causes for these deposits: post-spawning mortality; catastrophic mass mortality; predation (gastric mass or regurgitate);

stratigraphical - transportation and winnowing of the deposit accompanied by a low sedimentation rate (coeval lag, ancient lag or transported concentration); and re-sedimentation. They concluded that most deposits will have been formed by more than one of these factors.

Both post-spawning mortality and catastrophic mass mortality can be dismissed as causes in this case. Post-spawning accumulates are characterised by mono-specific populations entirely composed of adults. It has been demonstrated that in the populations of the Beeston Chalk an age range of individuals from juvenile to adult exists, maximum-sized specimens being relatively rare (Godwin, 1998a). The causes of catastrophic mass mortality are varied. Volcanic ash falls, a sudden onset of anoxia, salinity and temperature changes have been cited to explain marine mass mortality. These would affect the entire biota and no such events have been detected in the Beeston Chalk to date.

Winnowing and low-sedimentation rates (stratigraphical condensation) will have played a part in forming the accumulations of belemnite guards on the incipient hardgrounds. However, there is little evidence for mass transportation of the sediment or re-sedimentation. This suggests that predation was the main cause in this instance.

CYCLICITY AND THE CHANGING FACIES OF THE BEESTON CHALK

The link between orbital variations (Milankovitch orbital periodicity) and Cretaceous rhythmic bedding sequences is now accepted by many authorities (Barron *et al.*, 1985; Doyle *et al.*, 1994). Cyclic changes which appear in the rock record, however, do not fit a simple pattern. This is due to orbital variations in precession (oscillation about the axis), obliquity (tilt of the axis) and eccentricity (elliptic variation of the Earth's orbit) having different periods (21,000, 41,000 and 100,000 years respectively). Other factors such as solar variations, mantle cycles and galactic dust clouds may play a part too. This means that in some cycles the effects on solar insolation are cancelled out and in others they are enhanced (House, 1985; Laferriere *et al.*, 1987). Estimations for the length of Cretaceous cycles vary from 20,000 to 120,000 years or more (Park and Oglesby, 1991) and House (1985) even suggests macro-cycles of up to 450,000 years to explain the extreme temperatures in the Mesozoic.

The dominant periodicity in the Chalk appears to be 40-50,000 years (House, 1985; Arthur *et al.*, 1984). The Research on Cretaceous Cycles (R.O.C.C.) Group (1986) suggested that flint bands are caused by productivity cycles and omission surfaces by

scour cycles of the same order. This gives an alternative to Hancock's (1975) dating by average sedimentation rates (17.2m per million years for the Campanian) which yields a figure of c.1,000,000 years for the known sequence at Caistor. There are 19 such cycles in the section at Caistor which give a range of 760-950,000 years, which is comparable.

Barron *et al.*, (1985) proposed that cycles which caused variations in insolation affected ocean and atmosphere circulation. This may have led to changes in dominant wind direction and thus changes to ocean current patterns. This mechanism could explain facies changes in the Beeston Chalk. The model proposed by Barron *et al.*, (1985) sees oxygen in the water column increasing through each small cycle, with sedimentation rates slowing and water-energy levels decreasing. The upper part of the cycle sees increasing numbers of small encrusters colonising shell debris as conditions become more stable and more oxygenated. Then temperatures and biological productivity increase and oxygen levels fall once more.

Felder's model and meso-cyclicity in the Beeston Chalk

The Beeston Chalk between Flint Bands 7 and 12 displays a similarity to the model of cyclic deposition proposed by Felder (1986). He divided a cycle into five phases (A-E). Phase A begins on top of a hardground or omission surface. The benthos is dominated by sessile forms, sedimentation rates and water-energy levels are high, the grain-size is coarse and sea-level relatively low.

In phase B the benthos is dominated by burrowing echinoids. As sedimentation rates and energy levels decrease the grain-size fines upwards and sea-level rises. These trends continue into phase C where burrows and bivalves predominate. This is the level at which flints form during diagenesis. Phases D and E are dominated by vertical burrows, indicating shallower water (Gall, 1983). Water-energy levels increase in phase D, as the grain-size coarsens, but the sedimentation rate falls in phase E leading to the development of an omission surface.

These phases condense in shallower water and individual stages may be completely absent. For instance in the sequence above the hardground below Flint Band 7 (phase A) phase B appears to be absent and phase C is represented by Flint Band 7. Phases E and D are not evident in the section as the bioturbation could not be studied in detail.

Above this, the intermittent flint below the acme occurrence of *Echinocorys* (phase B) probably represents phase A of the next cycle. Phase C is represented by Flint

Band 8. Phases D and E are probably present but this part of the sequence is not well exposed. Similar partial sequences are found higher in the section.

Macro-cyclicity in the Beeston Chalk

A summary of the faunal and major sedimentological facies changes in the Beeston Chalk at Caistor St. Edmund is given in Figure 7 (based on data from Wood (1988), Peake and Hancock (1970), Pitchford (1991) and Christensen (1995) and this study). A full list of species can be found in Wood (1988) and Pitchford (1989). An interpretation of this information is given here in terms of apparent (Milankovitch) meso-cycles with periods of 20-100,000 years overprinted by macro-cycles which appear to be in the order of 250-300,000 years. It is proposed that 3 macro-cycles are present.

The proposed macro-cycles in the Beeston Chalk occurred during a "mega-cyclic" trend of global sea-level rise (or transgression). Therefore, transgressive tendencies are dominant within the macro-cycles. It is proposed here that they may reflect eustatic sea-level change. Transgressive pulses are indicative of global warming, with higher marine organic productivity, and a greater chance of anoxic-dysaerobic conditions prevailing. With regression, bottom currents increase and provide improved mixing in the water column and more stable substrates for a diverse benthonic fauna to develop upon.

Belemnites, being nektonic and unaffected by sea-floor oxygen levels, are not tied directly to this apparent macro-cyclicity. They are, however, connected indirectly through the food web, which controls their overall abundance. It is also proposed that their population structure, with either the *B. mucronata* or the *B. langei* group being dominant, is controlled by meso-cycles. The latter group dominated in transgressive phases and the former during regressions. The presence of abundant belemnites may indicate optimum conditions.

Macro-cycle 1

Regression. At the base of the sequence is the *Baculites* Bed. Rare ammonites are also found as far as Flint band 3. *Baculites*, given its straight, conical phragmacone and usually tear-shaped cross-section is thought to have had a nekto-benthic habit, in shallow environments down to 150-200m. Rarer species with an oval cross-section may have had a pseudo-planktonic life-style, drifting in surface waters down to 50m (Batt, 1989; 1991). Seilacher (1995) has recently suggested that ammonites functioned as "Cartesian divers"

Flint Bands

Flint band interval	0–1	1–2	2–3	3–4	4–5	5–6	6–7	7–8	8–9	9–10	10–11	11–12	12–13	13–14	14–15
Belemnite abundance (r - rare; c - common; a - abundant)		r	r	r	r	a	a	c	c	r	a	r	c	a	c
B. langei/B. mucronata group dominant (L/M)		L	L	?	?	M	M	L	L M	L	L M	L	L	L	L
Ammonite abundance - *Baculites/Neancyloceras*	a	r	r												
Brachiopod abundance (r - rare; c - common; a - abundant)	a	?	a	?	?	?	r	r	r	?	?	c	c	r	r
Brachiopod diversity (number of recorded species)	2		13	2			3	1	1			4	4	2	2
Brachiopods dominated by small forms (X)			X					X	X			X	X	X	X
Irregular echinoid abundance			c	?	?	c	c	c	a		a	a	?	c	?
Echinoids: large - L; small - S; thin-tested - T			T						L			S			
Regular echinoid debris present						X	X								
Crinoid abundance (r - rare; a - abundant)							r?						a		
Inoceramid "flood" - (X)							XXXXXXXXX					XXXXX			
Bivalve abundance (r - rare; c - common; a - abundant)	c	a	c	r	c	?	a	a	a	r	r	rc	a	?	?
Hexactinellid sponges present (X)	?	?	?	?	?	?	X	X	X	X	X	X	?	?	?
Small scleractinian coral absent (-); present (X)	?	?	?	?	?	?	X	X	X	X	-	-	?	?	?
Paramoudras, ring flints present (X)				X	X									X	
Incipient hardgrounds or omission surfaces present	X		X			X	X		X			X			
Evidence for dysaerobia (D); oxygenation (O)	O	O	O	D	D	O	O	D	D	D	D	O	D	D	D
Regressive - (R); Transgressive (T) tendencies	R	R	R	T	T	R	R	T	T	T	T	R	T	T	T
	< macro-cycle 1 >					< macro-cycle 2 >						<macro-cycle 3>			

Fig. 7. Faunal and environmental summary of the Beeston Chalk at Caistor St. Edmund. Information derived from Wood (1988), Peake and Hancock (1970), Christensen (1995) and this study. Data on micromorphic brachiopods not included.

and were capable of rapid vertical movements in the water column. Baculitids have a patchy distribution in the Upper Cretaceous of northwest Europe and North America. This suggested to Batt (1993) that they were limited by bottom-water dysaerobia. Their presence at the base of the sequence thus suggests well oxygenated waters, and it is here that trochid gastropods make their only appearance in the succession. Belemnites are rare in these lower beds.

The *Neoliothyrina* Bed occurs between Flint Bands 1 and 2 and is dominated by brachiopods. These are particularly abundant and diverse between Flint bands 2 and 3 in the *Orbirhynchia* Bed, but are predominantly small. Wood (1988) regards this as a transported assemblage as many of the specimens are disarticulated. *Echinocorys* from this bed are notably thin-tested, which is an indicator of dysaerobic conditions (Bottjer and Savdra, 1990). At this level ammonites make their last appearance in the Beeston Chalk at Caistor St. Edmund, although they are known to occur higher in the sequence elsewhere in Norfolk (Paul Whittlesea, pers. comm.). This may suggest the onset of dysaerobic bottom-water conditions or increased turbidity.

Transgression. Between Flint Bands 3 and 5 the preserved fauna is very sparse. Local paramoudras and ring flints occur. The former are associated with the "worm" burrow *Bathichnus paramoudrae*. Funnell (1986) suggested these may have been formed due to the activity of a tube worm, akin to those which inhabit mid-oceanic ridges today. If this was the case they may have had a hydrogen sulphide based metabolism. Given the paucity of the macrofauna it seems probable that conditions were oxygen deficient here. Flint Band 5 marks the end of the first macro-cycle.

Macro-cycle 2

Regression. Flint Bands 5 to 7 contain a pair of omission surfaces which must represent periods of bottom-water oxygenation and increased current activity. The benthonic community is very mature and diverse. Encrusting forms are common and regular echinoid test debris has been noted from this horizon as well as the presence of mosasaurs. Belemnites are abundant on these hardgrounds and seem to be dominated by the *B. mucronata* group.

Transgression. The inoceramid "flood" which begins just above the hardground below Flint Band 7 is indicative of depleted sea-floor oxygenation. Abernam (1994) cites *Inoceramus* as the main member of the epi-fauna in such contexts. It appears to have been an opportunistic coloniser of marginal dysaerobic environments, its dispersal being aided by pelagic larvae. This shelly chalk continues up to Flint Band 9 and the fauna is characterised by numerous belemnites (dominated by the *B. langei* group) and large specimens of *Echinocorys* (*Echinocorys* Bed, as detailed above).

Between Flint Bands 9 and 10 the chalk is generally unfossiliferous, although Christensen (1995) records the first appearance of two new species (*B. pauli* and *B. najdini*) which made the belemnite fauna more diverse. However, specimens are rare. The chalk remains phosphatic and this interval has been interpreted as a continuation of the transgressive phase of the macro-cycle. It appears to have been a "marine" desert which is more typical of the basinal chalks of southeast England.

The intermittent hardground above Flint Band 9 may reflect local scouring. The interval between Flint Bands 10 and 11 features numerous belemnites (dominated by the *B. langei* group) that appear to have been deposited as predator regurgitates. Flint Band 11 marks the termination of macro-cycle 2. The flint band, itself, is characterised by abundant juvenile echinoids. Some are grouped in "nests" of up to 20 individuals which have become incorporated into the flint (Wood, 1988).

Macro-cycle 3

Regression. A return to regressive conditions is seen in the hardground above Flint Band 11, which represents a return to more oxic bottom-water conditions. Brachiopods become more common and moderately diverse. "Giant" specimens of the *B. mucronata* group are common within or just above this surface. This phase is the shortest of the sequence. Such asymmetry in cycles has been noted by Laferriere *et al.*, (1987), and may have been caused by interference between differing orbital periodicities cancelling each other out, or locally due to erosion associated with sea-level rise.

Transgression. A second inoceramid "flood" occurs just below Flint Band 12 and continues to Flint Band 13, an interval which includes the *Austinocrinus* Bed. The fauna here is dominated by small brachiopods and the *B. langei* group appears to be dominant (Wood, 1988). Small pectinid bivalves and oysters also occur in this bed. Belemnites are

particularly abundant between Flint Bands 13 and 14 where paramoudra also occur. All these factors suggest dysaerobic conditions.

This interpretation suggests that transgressive tendencies accelerate and become more dominant through the three macro-cycles.

CONCLUSIONS

During macro-cycle 1 (Flint Bands 0-5) regressive tendencies are dominant and the fauna, in the lower part, is typical of an oxygenated off-shore chalk. Belemnites are rare during this interval but are dominated by the *B. langei* group. During the transgressive phase the sediment may have been intercepted by the oxygen-minimum zone (OMZ) as the benthonic macrofauna is largely excluded. Speciation of the belemnite fauna begins at this level with the appearance of the local species *Belemnitella* sp. 1.

At the start of macro-cycle 2 sea-level fell and shallow-water conditions probably encouraged migration from near-shore areas on the massif as adults of the *B. mucronata* group dominate the belemnite populations; however, the *B. langei* population appears to have been resident at this time. During the longer transgressive phase the OMZ may have intercepted the sea-bed for much of the interval because the macro-fauna is largely excluded except at certain discrete horizons, with the exception of *Inoceramus* whose pelagic larvae and tolerance to low-oxygen conditions are well-known (Abernam, 1994). The association noted by Wood (1988) of small (stunted?) brachiopods and *B. langei* might be explained by dysaerobia.

The communities which were studied in detail appear to have affinities with modern seamount assemblages (Rogers, 1994). *B. pauli* (a local species) *and B. najdini* whose origin is uncertain appear towards the end of the macro-cycle providing further evidence of endemism. Evidence for up-welling is suggested by the deposition of phosphatic chalk (Jarvis 1980a-c), and the presence of top predators. These suggest that the Caistor area was a rich feeding ground. Belemnites are rare in all other off-shore locations (Christensen, 1975).

Macro-cycle 3 sees deepening water levels (suggested by the presence of crinoids and the exclusion of corals); transgressive trends dominate, and dysaerobia appears to be a feature of the entire cycle. The *B. langei* group dominated the nekton, as in the transgressive phase of the other macro-cycles.

In conclusion, the distribution and speciation of *Belemnitella* in the Beeston Chalk at Caistor St. Edmunds appears to have been controlled by two factors: 1) the Mid-Campanian transgression and 2) the isolate palaeogeography/palaeoceanography of the site.

ACKNOWLEDGEMENTS

The author would like to thank Paul Whittlesea and Peter Murphy for their help during this study and Julian Andrews, Peter Doyle and Andrew Pitchford for their comments and advice on reorganising the original manuscript. Many thanks also go to Phillip Judge who drew the final versions of the diagrams.

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APPENDIX 1: Fossil Inventory

Hardground below Flint Band 7

Nekton

Belemnites: *B. mucronata* group (*B. minor I* Jeletzky, *B.* sp. 1 Christensen); *B. langei* group (*B. langei* Jeletzky).

Reptile: *Leiodon anceps* Owen (mosasaur)

Benthos

Bivalves: *Inoceramus balticus*? Bohm; Inoceramus spp.

Crinoids: *Austinocrinus bicoronatus* (Hagenow)

Echinoids: Regular and irregular echinoid spines and plate

Bryozoa: Small encrusting forms

Corals: Small, ring-shaped scleractinian corals

Sponges: Small, limonised and unidentifiable hexactinellid species

Echinocorys Bed

Nekton

Belemnites: *B. mucronata* group (*B. minor I* Jeletzky, *B.* sp. 1 Christensen); *B. langei* group (*B. langei* Jeletzky); *B. aff. americana* (Morton).

Fish: Impressions of fish scales

Benthos

Bivalves: *Inoceramus* spp.; *Acutostrea* cf. *incurva* Nilsson; fragments of unidentifiable oysters.

Brachiopods: *Cretirhynchia norvicensis* Pettitt

Echinoids: *Echinocorys aff. conoidea* (Goldfuss); *Galerites vulgaris* (Leske); *Cardiaster granulosus* (Goldfuss)

Bryozoa: Large free-lying and small encrusting forms

Corals: Small, ring-shaped scleractinian corals

Sponges: Hexactinellid, small, limonised, unidentifiable species

Between Flint Bands 9 and 10

Nekton

Belemnites: *B. mucronata* group (*B. minor* I Jeletzky, *B.* sp. 1 Christensen); *B. langei* group (*B. langei* Jeletzky)

Benthos

Bivalves: *Inoceramus* spp.; fragments of unidentifiable oysters.

Brachiopods: *Cretirhynchia norvicensis* Pettitt

Echinoids: *Echinocorys aff. conoidea* (Goldfuss).

Bryozoa: Small encrusting forms

Corals: Small, ring-shaped scleractinian corals

Sponges: Small, limonised and unidentifiable hexactinellid species

Between Flint Bands 10 to 12

Nekton

Belemnites: *B. mucronata* group (*B. minor I* Jeletzky, *B.* sp. 1 Christensen, *B. pauli* Christensen); *B. langei* group (*B. langei* Jeletzky, *B. najdini* Kongiel).

Benthos

Bivalves: *Inoceramus* spp; *Acutostrea* sp.

Brachiopods: *Cretirhynchia norvicensis* Pettitt; *Carneithyris carnea* (J. Sowerby); *Kingena* sp.

Echinoids: *Echinocorys* aff. *conoidea* (Goldfuss).

Sponges: Small, limonised and unidentifiable hexactinellid species

Trace fossils: *Thalassinoides* (crustacean burrow); *Zoophycos*

CONTENTS

The Geological Society of Norfolk exists to promote the study and understanding of geology, particularly in East Anglia, and holds monthly meetings throughout the year.

Visitors are welcome to attend meetings and may apply for membership of the society. For further details write to The Secretary, Geological Society of Norfolk, Castle Museum, Norwich NR1 3JU.

Copies of the Bulletin may be obtained from the Secretary at the address given above; it is issued free to members.

The illustration on the front cover is part of figure 5 from the third article by Godwin in this issue of the Bulletin. It shows belemnites from the *Echinocorys* Bed of the Beeston Chalk in Caistor St. Edmund quarry near Norwich. The two specimens on the right are *Belemnitella minor* I; the rest are *Belemnitella langei*. Note predator bite marks, borings and encrustations. Sketches are reproduced at c. x0.9 actual size.

ISSN0143-9286

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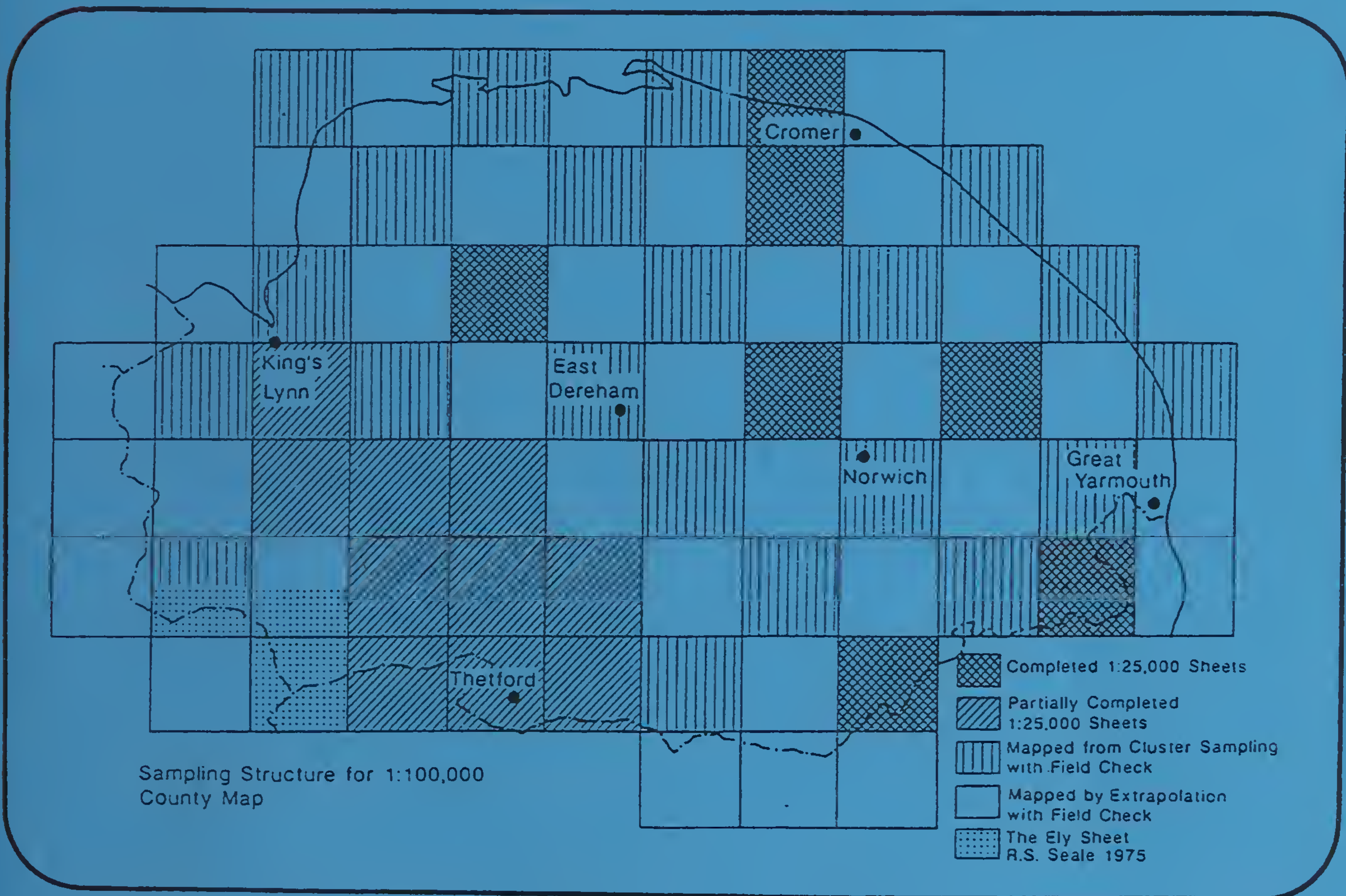
BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

(FOR ARTICLES ON THE GEOLOGY OF EAST ANGLIA)

NO.48 1998



PUBLISHED 1999

CONTENTS INCLUDE

Bone & pyrite diagenesis

Norfolk soil maps

March Gravels

BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

No. 48 (for 1998) Published 1999

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EDITORIAL

Bulletin No. 48 contains three very different papers on various aspects of geology in Norfolk. The papers by Richard West and Bill Corbett, on the March Gravels and Norfolk Soils respectively, return to themes that the Bulletin has published in the past. These are fascinating and welcome updates. The paper by Gordon Turner-Walker takes another look at the diagenesis and chemistry of the West Runton Freshwater Bed, this time from the perspective of bone preservation. The results seem to bear out earlier work on the geochemistry of the deposit itself published in Bulletin No. 45.

The appearance of Bulletin 48 keeps the publication schedule more or less up to date. Copy for Bulletin 49 is in hand and will be published shortly, which means that we should go into the year 2000 right up to date. As usual I welcome the continued submission of papers on all aspects of East Anglian geology.

INSTRUCTIONS TO AUTHORS

If possible, contributors should submit manuscripts as word-processor print out accompanied by a disk copy. We can handle most word-processing formats although PC Word, WordPerfect or ASCII files are preferred. In addition we accept typewritten copy and will consider legible handwritten material.

It is important that the style of the paper, in terms of overall format, capitalisation, punctuation, etc. conforms as strictly as possible to that used in Vol. 41 of the Bulletin. Titles and first order headings should be capitalised, centred and in bold print. Second order headings should be centred, bold and lower case. Text should be 1½ line spaced. All measurements should be given in metric units.

References should be arranged alphabetically in the following style.

BALSON, P.S. & CAMERON, T.T.J. 1985. Quaternary mapping offshore East Anglia. *Modern Geology*, **9**, 221-239.

STEERS, J.A. 1960. Physiography and evolution: the physiography and evolution of Scolt Head Island. In: Steers, J.D. (ed.) *Scolt Head Island* (2nd ed.), 12-66, Heffer, Cambridge.

BLACK, R.M. 1988. *The Elements of Palaeontology*. 2nd Ed., Cambridge University Press, Cambridge. 404pp.

Illustrations should be drawn with thin dense black ink lines. Thick lines, close stipple or patches of solid black should be avoided as these spread in printing. Original illustrations should, before reproduction, be not more than **175mm by 255mm**. Full use should be made of the first (horizontal) dimension which corresponds to the width of print on the page, but the second (vertical) dimension is an upper limit only. Half tone photographic plates are acceptable when their use is warranted by the subject matter, provided the originals exhibit good contrast.

The editors welcome original research papers, notes or comments, and review articles relevant to the geology of **East Anglia** as a whole, and do not restrict consideration to articles covering Norfolk alone. All papers are independently refereed by at least one reviewer.

Pyrite and Bone Diagenesis in Terrestrial Sediments: Evidence from The West Runton Freshwater Bed

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ABSTRACT

Recent research into the carbon and sulphur geochemistry of the organic-rich, Pleistocene sediments of the West Runton Freshwater Bed (Norfolk, U.K.) have demonstrated how groundwater effects can bring about post-diagenetic remobilisation of pyritic sulphur and how this, in turn, can influence subsequent use of bulk C/S ratios as a tool to distinguish freshwater from marine sediments. Microscopic examination of fossil bones from the Freshwater Bed illustrates some of the initial stages of pyrite diagenesis and the transition from finely divided framboidal pyrite to more massive forms. In addition, the state of preservation of histological structures in the fossil bones reflect the chemical environment prevailing in the sediments in which the bones became buried. This burial environment has contributed to the remarkably good state of preservation of fossils from West Runton.

INTRODUCTION

Following the discovery of two fossil elephant bones eroding from the cliffs at West Runton, Norfolk in December 1990, two rescue excavations were undertaken to recover others bones visible in the section to prevent them being destroyed or dispersed by winter storm erosion. These excavations were undertaken in January 1992 and from the bones recovered the elephant was identified as an ancestral mammoth, *Mammuthus trogontherii* (Driscoll and Stuart, 1992; Stuart, 1993, 1997). In October 1995 a major excavation to locate the remaining bones resulted in the retrieval of almost the complete skeleton. The sediments from which the fossil elephant was recovered, the West Runton Freshwater Bed (WRFWB), forms part of the Cromer Forest Bed Formation. It is the type site for the

Bull. geol. Soc. Norfolk (for 1998) **48**, 3-26 (Published 1999)

Cromerian interglacial of the Pleistocene stage of the Quaternary, and the sediments are approximately 600,000 years old. Since the early 19th century numerous fossil remains of elephants, rhinoceros, deer and other mammals have been collected from the Cromer Forest Bed. However, the discovery of a complete skeleton was of unique. The WRFWB (found between grid references TG 186431 and TG 189431; Fig. 1) comprises a layer of organic-rich detrital silt, approximately one metre thick, exposed at beach level and overlain by a sequence of marine and freshwater sands and silts up to 19 m thick that make up the bulk of the cliff. The 1995 excavation was undertaken by the Norfolk Archaeological Unit using accepted archaeological techniques and recording procedures. This represented perhaps the first occasion in Britain when a professional archaeological team had been contracted to work on a non-human fossil site (Ashwin and Stuart, 1996). Before the excavation could proceed, it was necessary to remove, by dragline, approximately 1330 cubic metres of sand and gravels from the cliffs overlying the proposed area of excavation, an area 15 m x 5 m which lay beneath the cliffs, immediately adjacent to the rescue excavation of 1992. The removal of the overburden not only enabled the later safe excavation and recovery of the remaining skeleton, but also allowed access to a large body of freshly exposed, clean and unweathered Freshwater Bed. Many Quaternary researchers took advantage of the opportunity to sample this pristine material, removing samples for pollen analysis, sedimentological and dating studies. In addition, over 10 tonnes of sediment samples were removed from site for later sieving to recover small vertebrate, insect and plant fossils. Preservation of fossil material is exceptionally good in the WRFWB which contains remains of beetles, small amphibians and fishes, mollusc shells and recognisable fragments of woody plant material. In particular, mammal bones from West Runton exhibit a remarkable degree of preservation and frequently preserve surface features and internal structures in microscopic detail.

A subject that has attracted considerable study is the palaeosalinity of the WRFWB. Previously, the ratio of organic carbon to pyritic sulphur had been used to distinguish between marine and freshwater sediments in the geological record (Berner and Raiswell, 1983, 1984). The application of this method to the WRFWB showed that in the upper and middle parts of the bed, pyrite sulphur had been oxidised and remobilised - probably by groundwater (Hannam *et al.*, 1996; Bottrell *et al.*, 1998). However, in the lower part of the WRFWB, the sediment is much better preserved and C/S ratios suggest

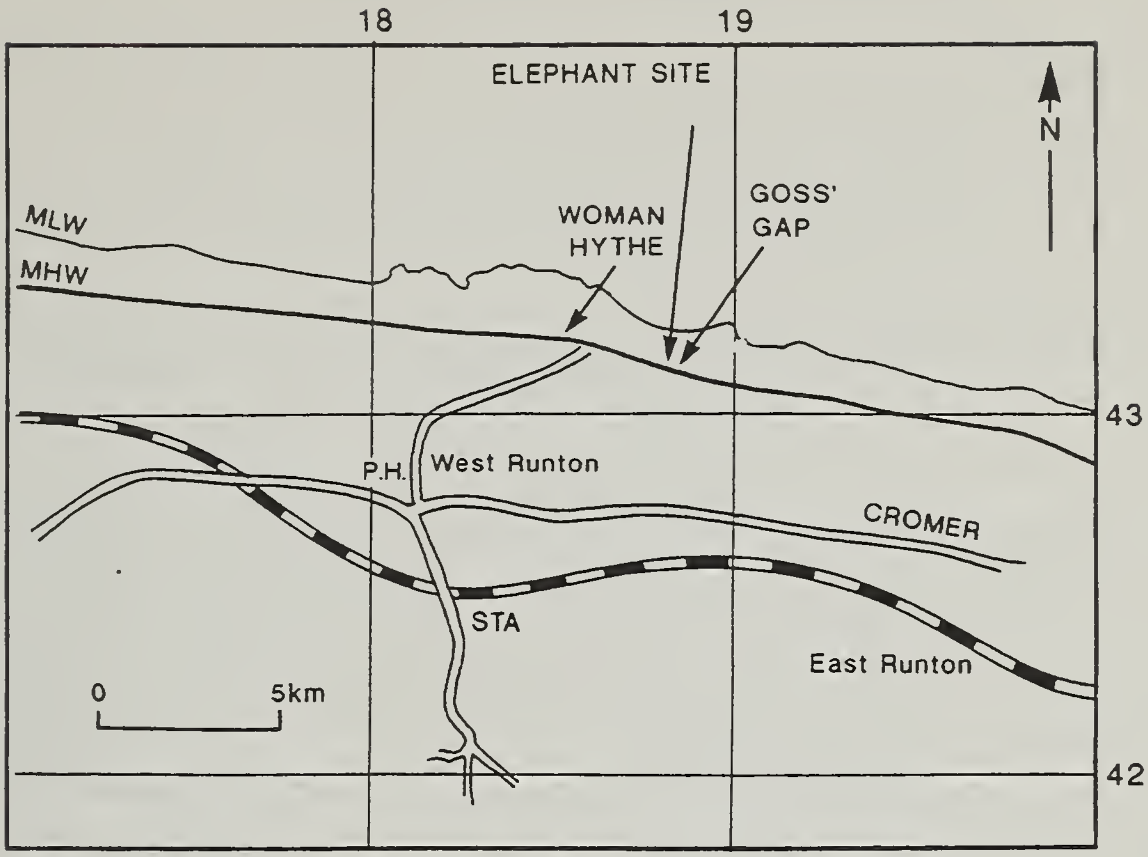

Fig. 1. Location map of the WRFWB elephant excavation site at West Runton (Ordnance Survey grid lines); after Hannam *et al.* (1996).

a freshwater depositional environment. Faunal and pollen assemblages suggest the deposit was formed in a freshwater channel flanked by deciduous woodland.

Pyrite has been identified in scanning electron microscope (SEM) images of skeletal material recovered during the 1995 excavations at West Runton and in combination with energy dispersive spectrometry (SEM-EDX) some of the early stages of pyrite diagenesis have been distinguished. Furthermore, in the bones examined to date, the absence of those microscopic features that are characteristic of the early stages of bone diagenesis in normal soils, suggests that the elephant skeleton underwent rapid burial within a relatively short time of the animal's death. Work on the taphonomy of the elephant carcass, involving analysis of the spatial distribution of skeletal elements and trace marks on the bones' surfaces is currently being undertaken at Norwich Castle Museum.

PYRITE IN TERRESTRIAL SEDIMENTS

One of the most important groups of micro-organisms responsible for the mineralization of organic carbon in sediments are the sulphate-reducing bacteria (SRB). The sulphate-reducing bacteria form a very diverse group and can metabolise a wide range of organic compounds, although they can only utilise low-molecular-weight compounds such as simple organic acids (lactate, acetate and propionate) which in turn represent the fermentation products from anaerobic bacterial degradation of more complex substrates. The SRB are widely distributed in the environment but are most common in anoxic sediments. They can, however, tolerate long periods of exposure to oxygen and as a group have adapted to almost every environment, covering a temperature range of -5 to 75 degrees Celsius, pH from 5 to 9.5 and almost any pressure or salinity. As a by-product of the reduction of sulphate ions, SRB release dissolved sulphide into the environment where it reacts with metal ions and detrital minerals to form highly insoluble sulphides, including pyrite.

In contrast to the main focus on marine sediments, much less work has been undertaken on the distribution of SRB and pyrite diagenesis in terrestrial sediments, partly as a result of the comparatively small volumes of sediment concerned (compared with marine sediments) and partly because dissolved sulphate concentrations in freshwaters are typically 1% of marine sulphate concentrations (Berner and Raiswell, 1984). Nevertheless, the World's peat bogs and marshlands are an important sink for organic carbon and SRB play an important role in the diagenesis of organic sulphur in anoxic bogs and peatlands (Brown, 1986; Brown and MacQueen, 1985; Novak *et al.*, 1994). Another environment in which SRB can flourish (albeit insignificant on a global scale) is in archaeological deposits, particularly deep cultural levels on urban sites. Here, a combination of relatively high soil water content, rapid accumulation of organic matter and other detritus associated with urban societies, and the rapid build up of sediments as towns develop and grow can lead to anoxic environments with levels of sulphate and dissolved metal salts far higher than those found in other terrestrial deposits. Pyrite, and framboidal pyrite in particular, have been identified during microscopic examination of archaeological finds from anoxic terrestrial "waterlogged" deposits rather than in the sediments themselves (which are rarely subject to analysis). Pyrite has been found in ancient organic materials such as wood (Watson, 1981) and bones (Day and Molleson, 1973; Parker and Toots, 1974; Ellam,

1985; Garland, 1987; Turner-Walker, 1993). It has also been found in close association with inorganic substrates such as pottery (Green, 1992) and corroding iron objects (Fell and Ward in press). Mixed iron and copper sulphides occasionally form gold coloured "crusts" on copper alloy artefacts from these deposits, leading sometimes to their confusion with deliberate pseudo-gilding (Duncan and Ganiaris, 1987).

MICROSCOPIC AND CHEMICAL ANALYSIS OF MATERIAL FROM WEST RUNTON

Materials and methods

Samples of bone and ivory from the 1995 excavation were selected for SEM examination and for chemical analysis. The samples comprised some fragments of ivory and skull found early in the course of the excavation, approximately 30 cm below the top of the WRFWB. These skull and ivory fragments were not from the main elephant skeleton but from another, much more fragmentary, individual. In addition to the bone and ivory found close to the top of the bed, some small fragments of an elephant rib found approximately 100 cm below the top of the WRFWB were analysed for carbon, hydrogen and nitrogen content, although these samples were not examined using SEM. Small samples of sediment and minerals removed from the surfaces of the bones were also examined by X-ray diffraction (XRD).

Prior to SEM examination, the fragments of bone and ivory (approximately 2-3 cm^3) were dried over silica gel and embedded in a transparent epoxy resin (Araldite AY103/HY991). The cured blocks were then polished to an optically flat surface using silicon carbide papers followed by a fine, aluminium oxide polishing compound. Before imaging, the samples were once again thoroughly desiccated over silica gel, then carbon coated to provide an electrically conductive surface. The samples were imaged in backscattered electron (BSE) mode using a LEO Stereoscan 440i fitted with a Link GEM X-ray detector having a light element window.

Small (approximately 1mg) samples for XRD analysis were ground in an agate mortar and spread onto a flat, single-crystal silicon sample holder. Spectra were collected using a Phillips PW 1840 diffractometer utilising cobalt K_α radiation and a solid state silicon detector. A search-match computer program was used to identify components of the spectra by comparison with standards in the powder diffraction file.

Samples for carbon, hydrogen and nitrogen analyses were run on a Carlo Erba 1106 Elemental Analyser and weighed using a Mettler MT 5 microbalance.

Results

Results of the carbon, hydrogen and nitrogen analyses (Table 1) show that the fossil material has lost a considerable proportion of its original organic content (i.e. the structural protein, collagen) and that the skull from the top of the bed has lost proportionally more than the elephant rib found lower in the deposit. Attempts to measure the collagen content of a sample taken from the elephant's pelvis (found in 1990) using weight loss following hydrazinolysis (Turner-Walker 1993), were unable to confirm the presence of any collagen. This is entirely in keeping with current models of collagen diagenesis (Collins *et al.* 1995) which suggest that the collagen in mineralised tissues buried at these latitudes should not survive in excess of 500,000 years. Furthermore, the relative proportions of carbon, hydrogen and nitrogen for all the fossil samples are substantially different to those in modern bone, in that the fossil material is proportionately richer in carbon and depleted in nitrogen. These observations may reflect the presence of humic compounds, or as the protein is depleted, that other, non-collagenous resistant macromolecules in the bone (for example lipids) skew the carbon/nitrogen ratio.

Table 1: Carbon, hydrogen and nitrogen analyses of bone and ivory (modern material for comparison).

Sample (Find No.)	Depth (cm)	wt%C	wt%H	wt%N
Upper skull (17.1)	c. 30	2.81	0.28	0.32
Upper skull (17.2)	c. 30	3.24	0.29	0.26
Upper tusk (9.1)	c. 30	2.97	0.16	0.25
Upper tusk (9.2)	c. 30	3.32	0.26	0.36
Elephant rib (52.1)	c. 100	4.16	0.31	0.84
Elephant rib (52.2)	c. 100	4.66	0.40	1.00
Fresh bone[1]	-	20.94	3.40	6.50
Elephant ivory[2]	-	11.7	1.9	3.6

[1] Values for modern sheep bone (Turner-Walker 1993)

[2] approx. values for modern elephant ivory after Krzyskowska (1990)

Electron micrographs (SEM-BSE) of polished sections of the bone and ivory samples collected from the top of the WRFWB are shown in Figures 2-5. SEM images using backscattered electrons carry information on the atomic number density of different regions of the sample, such that the intensity is related to the mean density of the sample under the electron beam. Denser areas of the sample scatter more electrons back towards the detector and consequently appear brighter in the SEM image. Figure 2a shows a low magnification view of a skull fragment which illustrates the natural porosity of the bone. The embedding resin, which has a very low atomic number density, appears black in the images compared to the bone which is largely composed of bioapatite, a stoichiometrically imperfect carbonate-containing, hydroxyapatite analogue ($Ca_{10}(PO_4)_6.(OH)_2$). The image readily demonstrates the excellent preservation of the bone with the normal micromorphology of (dead) bone tissue readily visible, with little evidence of the degraded structures frequently seen in diagenetically altered bones. The very bright areas lining the pores represent diagenetic pyrite, predominantly in the form of discrete framboids. These can be seen more clearly at higher magnification in Figure 2b which covers the area of the top left corner of Figure 2a. This image shows numerous circular aggregates of small crystals (cross-sections through spherical pyrite framboids) in addition to an area in which several framboids have coalesced to form a mass of closely packed crystals. It is noteworthy that, apart from the diagenetic pyrite, the natural porosity of the bones from West Runton remain void and have not been filled by infiltrating sediment. In addition to the natural, physiological porosity, which has dimensions that range from 40 μm to several hundred microns, there is evidence for some additional porosity resulting from post-mortem microbial degradation processes (Fig. 2c). This diagenetic porosity, commonly referred to as "tunnelling" (Hackett, 1981) is restricted to the outer surface of the bone where it was in contact with the surrounding sediment, and to the bone immediately surrounding the natural pores (Fig. 2c). This tunnelling has much smaller and more consistent dimensions, typically 10 μm in diameter. Present evidence indicates that tunnelling in archaeological and fossil bones results from bacterial action in which the bone is demineralised in order to expose the proteinaceous collagen matrix which makes up 23% of the bone by weight. This highly attractive food source is otherwise unavailable for utilisation by micro-organisms since it is effectively protected by its intimate association with the cryptocrystalline bone mineral apatite. Figure 3 shows a high magnification view of framboidal pyrite lying within a pore in the interior of the bone

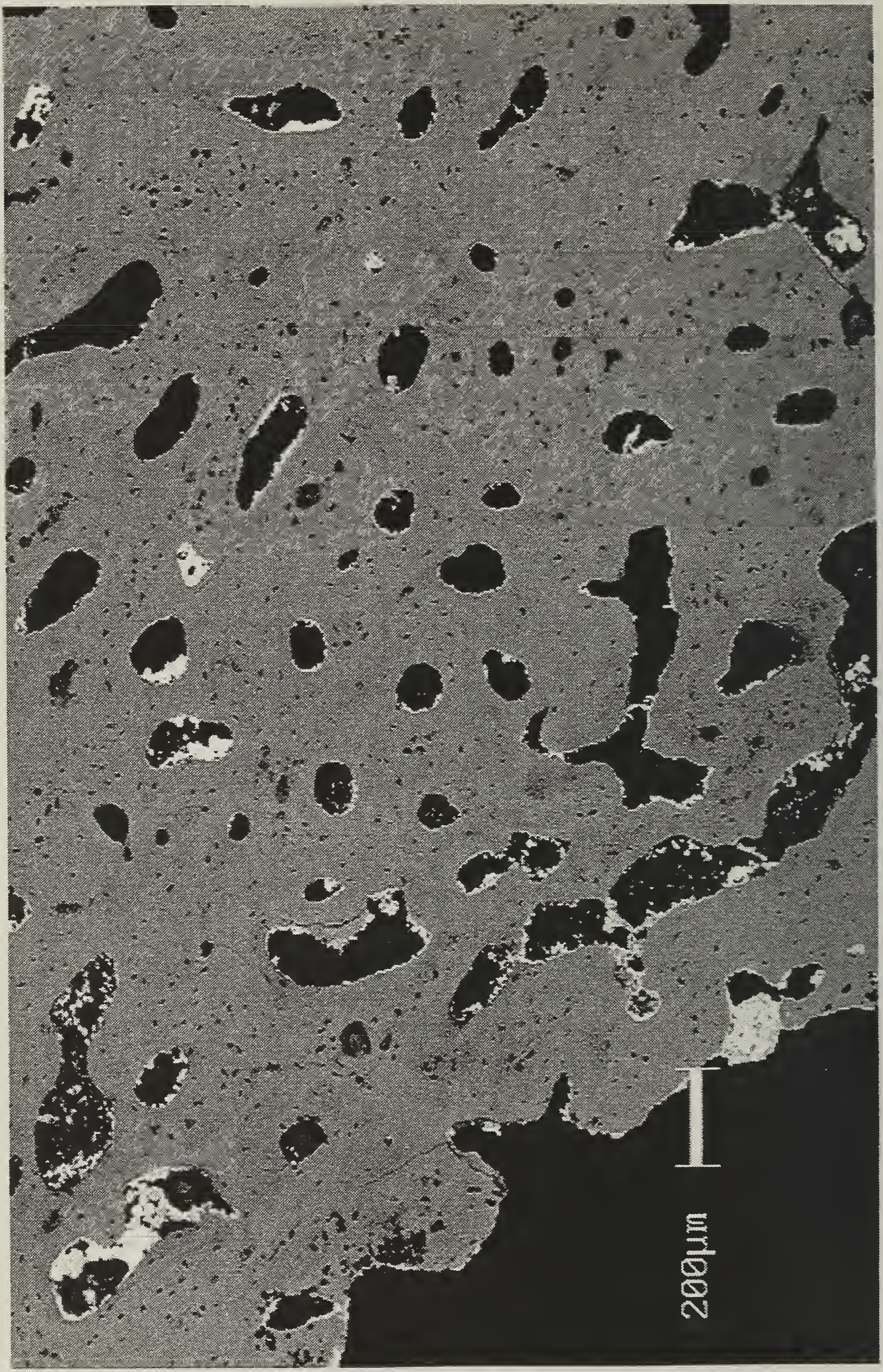

Fig. 2a. SEM-BSE photomicrograph of skull fragment from top of WRFWB.

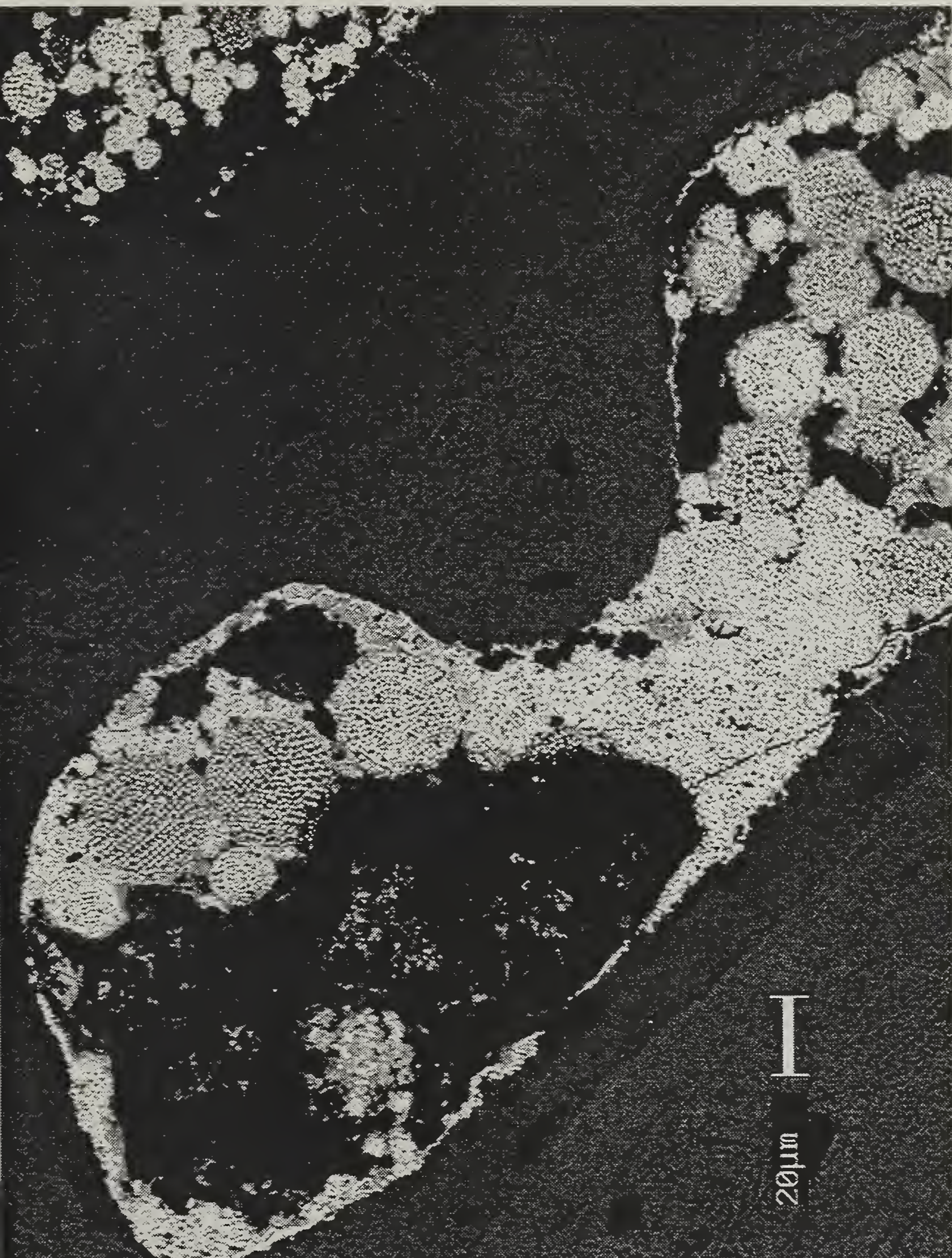

Fig. 2b. SEM-BSE photomicrograph of detail of Figure 2a showing numerous pyrite framboids.

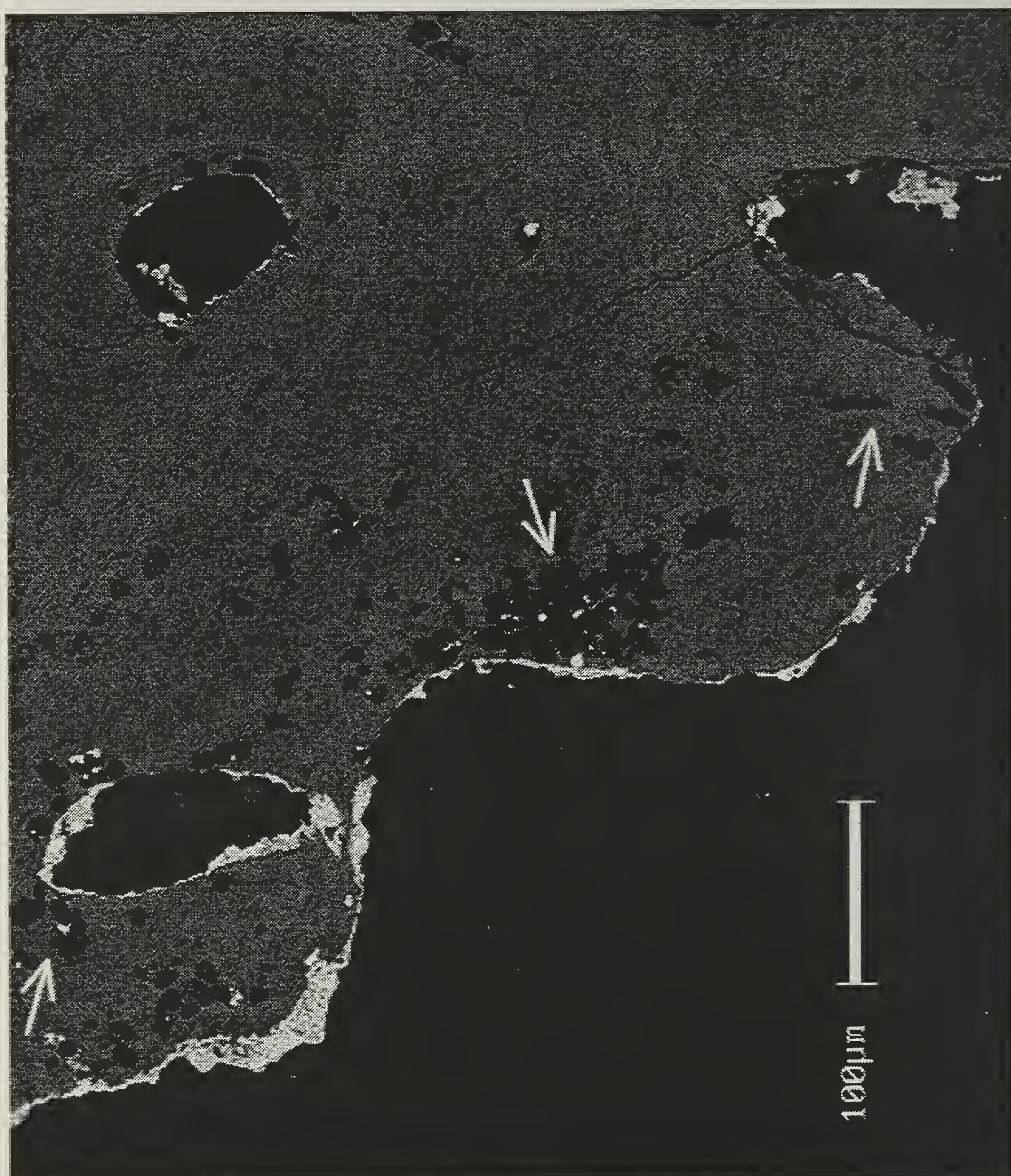

Fig. 2c. SEM-BSE photomicrograph showing diagenetic tunnelling resulting from post-mortem microbial degradation.

fragment (not visible in Figure 2a) and on which EDX analysis was undertaken. Analysis of a single grain in the centre of the field confirmed that the individual grains were iron sulphide with a composition close to FeS_2 (Table 2). The largest grains visible in this image have dimensions of the order of 1μm. Analysis of the arrowed feature (diameter approximately 5 μm) of lower density material in the upper centre of Figure 3 showed that this material contained iron oxide, with only traces of sulphur and phosphorus (Table 2).

Backscatter images of a longitudinal cross-section of tusk showed a similar pattern of excellent preservation, although ivory is much less porous than bone being composed of closely aligned "dentinal tubules". In Figure 4a these are highlighted by being infilled by dense diagenetic pyrite. To the right of the image a layer of dense, massive pyrite overlays the surface of the tusk and this, in turn, binds a thin layer of sediment to the outer surface. Towards the interior of the tusk, pyrite not only fills the dentinal tubules but also fills longitudinal cracks in the ivory (Fig. 4b). EDX analysis of the material filling the cracks indicates it is iron pyrites (Table 2). The ivory itself was also subjected to EDX which gave results consistent with bioapatite (Table 2). The missing mass of 3.46 weight percent was also in good agreement with the carbon, hydrogen and nitrogen analyses (totalling 3.38% and 3.94% for the two samples shown in Table 1). The relatively high fluorine content (2.44 wt.%) is also noteworthy, fluorine being used to date fossils before the advent of radiocarbon dating. Traces of fluoride in groundwaters undergo substitution reaction with bioapatite resulting in a gradual transformation to fluorapatites over geological timescales.

Samples subjected to X-ray diffraction analysis included soil scraped from the surface of the a rhinoceros bone recovered from the base of the WRFWB; white efflorescent material taken from the rhinoceros bone and a yellow, lustrous mineral scraped from the surface of the rib and assumed to be pyrite. A bone fragment from the rhinoceros was also analysed. Results of the XRD analyses are summarised in Table 3. The distinction between major and minor components was made purely qualitatively on the grounds of relative peak heights in the XRD spectra. All samples were taken six months after the date of the excavation so it is inevitable that some of the mineral species identified represent oxidation or dehydration products of minerals or solutions present in the original material.

Table 2: Results of EDX analyses on samples from fossil bone and ivory.

Sample	**Pyrite framboid in skull**		**Circular feature in skull**[1]		**Mineral in crack in ivory**		**Ivory**	
Element	*wt %*	*atom %*	*wt%*	*atom %*	*wt %*	*atom %*	*wt%*	*atom%*
O	12.98	25.96	28.84	62.18	3.78	8.99	44.38	65.04
F	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	2.44	3.01
Na	0.25	0.35	0.50	0.71	0.17	0.28	0.46	0.47
Si	0.26	0.29	0.57	0.70	0.09	0.12	1.17	0.98
P	0.23	0.23	2.18	2.43	0.02	0.02	14.68	11.11
S	46.73	46.61	1.76	1.89	50.65	60.08	0.32	0.24
K	0.02	0.02	0.07	0.06	0.06	0.06	n.d.	n.d.
Ca	0.43	0.34	0.91	0.78	0.12	0.11	31.87	18.64
Mn	0.08	0.04	0.14	0.09	0.12	0.08	0.09	0.04
Fe	45.82	26.24	50.44	31.25	44.43	30.26	1.15	0.48
Total	**106.64**	**100.00**	**106.64**	**100.00**	**99.43**	**100.00**	**96.54**	**100.00**

[1] Arrowed circular feature of low backscatter intensity visible in Figure 4.
wt% = weight %; atom % = atomic %

Table 3: Results of XRD analyses (qualitative).

Sample No.	**Source**	**Major components**	**Minor components**
N37.RD	efflorescence	gypsum, quartz	-
N38.RD	soil	quartz	pyrite, gypsum, albite?
N39.RD	soil	quartz	pyrite, gypsum, melanterite?
N40.RD	yellow, lustrous mineral	pyrite, quartz	-
N41.RD	bone	hydroxyapatite	-

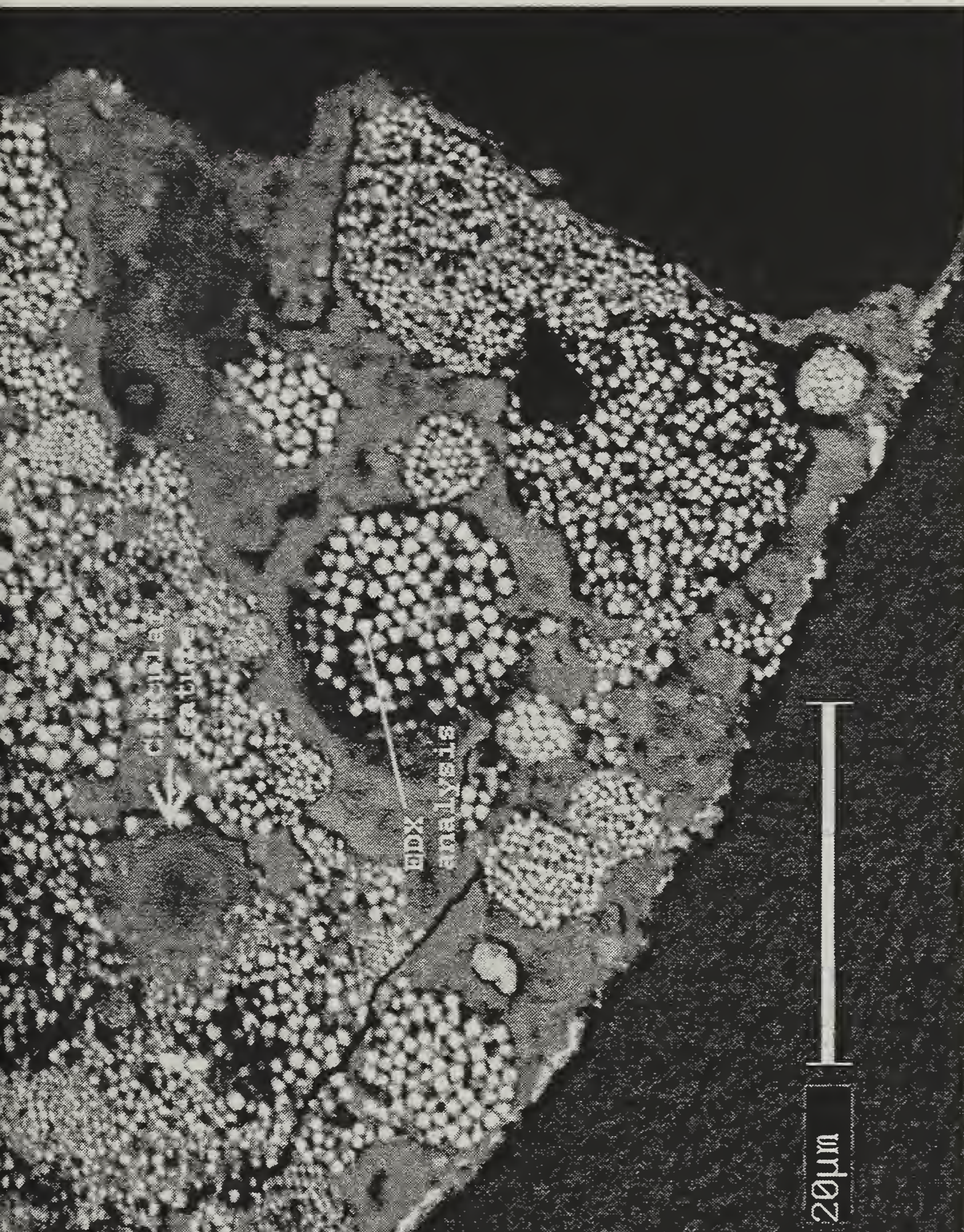

Fig. 3. SEM-BSE photomicrograph of framboidal pyrite on which EDX analysis was undertaken.

DISCUSSION

The SEM, XRD and carbon, hydrogen and nitrogen analyses each confirm the general, visual appearance of the fossil material as being in an exceptional state of preservation, albeit depleted of its collagen component. The analyses also confirm the presence of pyrite within natural pores in the bone and ivory and encrusting the surface of the ivory. Although most of the pyrite is present in the form of discrete framboids, there is some evidence in the bone for the coalescence of individual framboids into a more solid mass and the ivory quite clearly has had post-depositional cracks filled by massive pyrite. The exact mechanisms of framboid formation are extremely complex and outside the scope of this study, however, there is general agreement that dissolved sulphide resulting from the metabolism of SRB reacts with iron either in the form of aqueous Fe^{2+} in the groundwater or detrital iron minerals in the surrounding sediment (Vaughan and Lennie, 1991). The reaction between dissolved sulphide and iron species has been proposed to pass through a number of intermediate stages including makinawite ($Fe_{1+x}S$ where x = 0-0.2) and greigite (Fe_3S_4) ultimately leading to pyrite which represents a thermodynamically stable iron sulphide. Marcasite (FeS_2; S-deficient) is an end-product polymorph favoured at low pH. All of the above sulphides have been identified in archaeological objects from “waterlogged” archaeological deposits (Fell and Ward, in press). Since the SEM images of the skeletal material from the WRFWB show that the pores are effectively void except for the presence of pyrite, the evidence suggests that in the bones at least, the sulphide must be reacting with dissolved iron species rather than solid particles as may be the case in the surrounding sediment. Both field and laboratory studies (Wilkin and Barnes, 1996, 1997) have confirmed that pyrite framboids can form in suspension from dissolved species and that a framboidal texture is favoured when iron monosulphides react rapidly to form pyrite (Wilkin and Barnes, 1996). Other research on pyrite formation in freshwater systems in the Netherlands has found pyrite formation associated with organic matter where it was thought to form in “microsites”. Here, conversely, the authors attribute the framboidal texture to slow rates of pyritisation (Marnette *et al.*, 1993).

The arrowed feature in Figure 4 has a composition consistent with an iron oxide with the formula FeOOH, possibly goethite or lepidocrocite. Its shape suggests that this may be a framboid that has undergone subsequent oxidation, although it is unclear whether this took place before or after the excavation of the bone. Similar low density

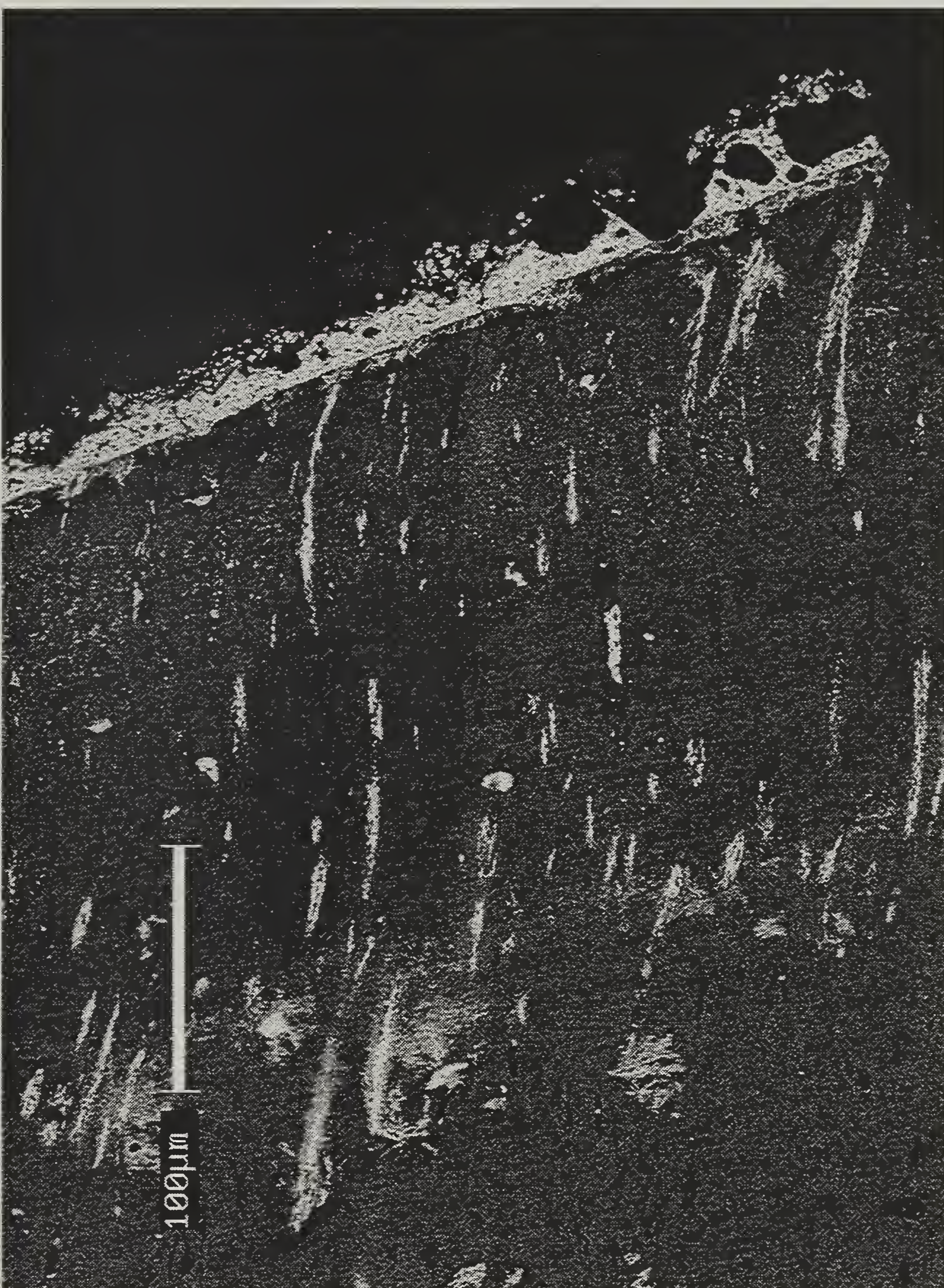

Fig. 4a. SEM-BSE photomicrograph of ivory showing dentinal tubules.

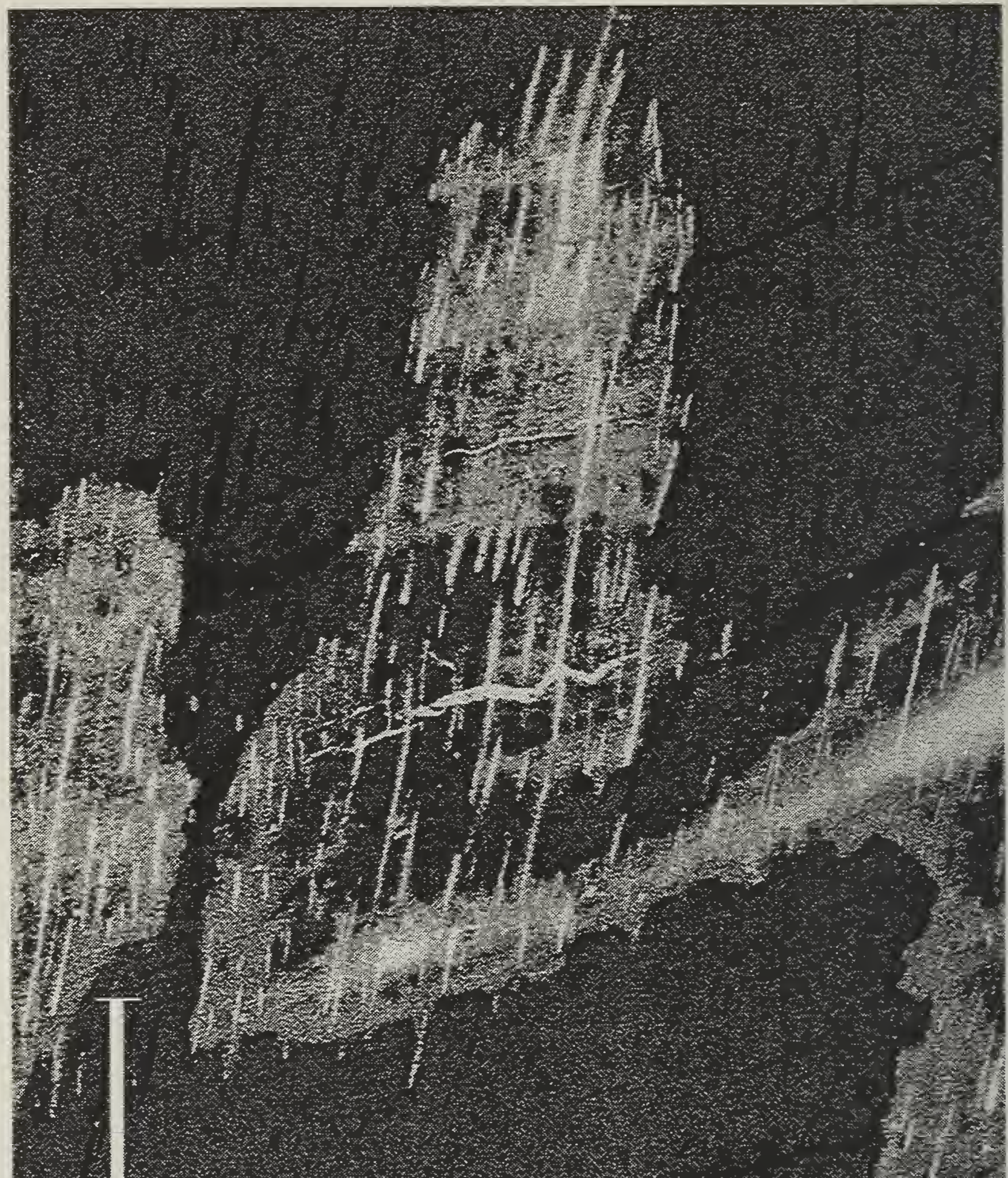

Fig. 4b. SEM-BSE photomicrograph of ivory showing pyrite filling dentinal tubules and post-depositional cracks.

features have been seen in close association with pyrite framboids in SEM-BSE images of ancient bones from other anoxic contexts. Elemental dot-mapping of these structures showed them to contain calcium and phosphorus but no sulphur (Turner-Walker, 1993). Electron micrographs of the bones from the WRFWB and from another Pleistocene deposit, the Ipswichian sediments exposed at Shropham (grid reference TM 000940), Norfolk, suggest that with increasing pyritisation, framboids may be consolidated into a solid mass of pyrite (Fig. 5). In this image, a large pore has been totally filled by massive pyrite deposits but closer examination reveals the remnants of several framboids embedded within the main body of compact pyrite. This change from framboidal to massive pyrite formation may reflect a fall in the rate of pyritisation.

The bone from the WRFWB is very well preserved histologically despite having lost all of its original organic matter. Current models of bone diagenesis recognise two, effectively independent, degradation pathways for the collagen component. One mechanism, which predominates in aerated soils, involves the action of aerobic bacteria which cause local demineralisation of the bone in advance of enzymatic degradation of the exposed collagen matrix. Acting in discrete zones these, as yet unidentified organisms, increase the porosity of the bone at a sub-micron level, leading ultimately to the creation of the larger channels identified by some researches as tunnels. The second degradation pathway is purely chemical (abiotic) and involves the gradual hydrolysis of the mineralised collagen and subsequent leaching of the resulting peptide fragments (Collins *et al.*, 1995). This is essentially a time-temperature dependant phenomenon with the degree of collagen loss determined by the antiquity of the specimen and the latitude and depth of its burial site. Only in exceptional circumstances where liquid water is excluded, for example in mummified or frozen remains, does the collagen component of bone remain intact (the exception to this rule appears to be some of the bones from Shropham in which over 90% of the original collagen remains after 120,000 years of burial).

The bones from the WRFWB have experienced the latter second degradation pathway, *i.e.* loss of collagen by hydrolysis and with negligible bacterial degradation. Furthermore, the XRD spectra of the bone sample showed the characteristic broad peaks of poorly crystalline hydroxyapatite. The tunnelling associated with microbial degradation of bone often leads to an increase in the average crystallinity, identifiable in XRD spectra as a distinct sharpening of the main diffraction peaks. The absence of an appreciable increase in the crystallinity of the West Runton bone is demonstrated in Figure 6 which

Fig. 5. SEM-BSE photomicrograph of a bison bone from the Pleistocene site at Shropham, Norfolk. Note the coalescence of several framboids into a mass of dense pyrite.

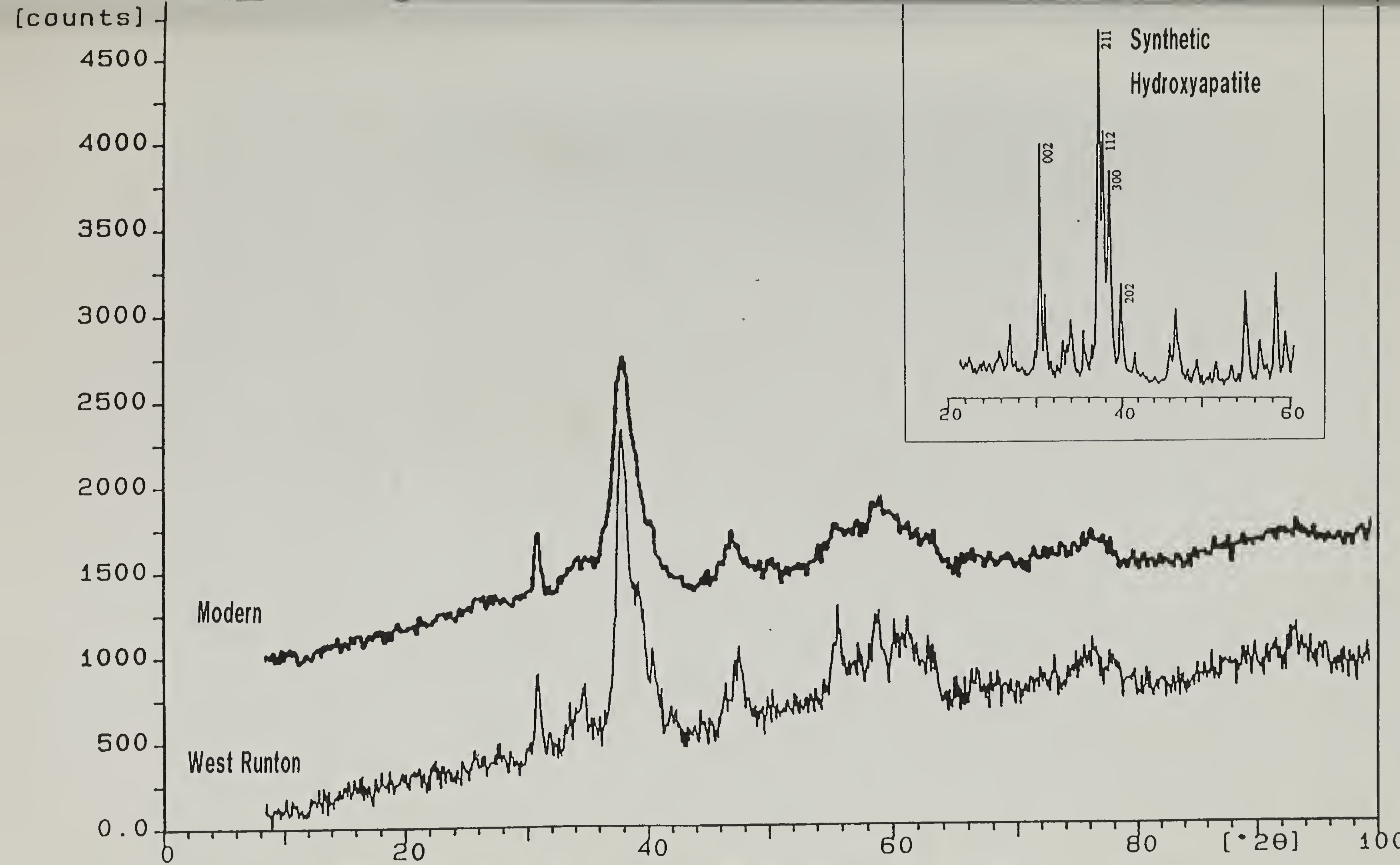

Fig. 6. Comparison of XRD spectra from modern animal bone (upper trace) with bone from towards the base of the WRFWB (lower trace). Compare the broad peaks around 38° 2θ with the three resolved peaks in synthetic hydroxyapatite (inset).

compares a bone from the base of the bed with modern animal bone. In well crystalline synthetic hydroxyapatite the main peak can be resolved into its three components, the 211, 112 and 300 peaks. In fresh and recent bone specimens the individual peaks cannot be resolved but appear as a broad, asymmetric peak as seen in the upper trace of Figure 6 at around 38° 2θ.

West Runton is not the only site to have produced bones in an excellent state of preservation and which also contain framboidal pyrite. Recent work on human bones from a medieval inhumation cemetery in Trondheim, Norway has shown that, here too, excellent histological preservation correlates with the presence of framboids in the natural pore spaces. In contrast to the bones from the WRFWB, the human bones from Trondheim still retain a considerable fraction of their original collagen content, having lain in the ground for barely 400-800 years. That such geologically young bone should contain pyrite is not surprising. The smallest pores of effectively fresh, modern bone have been shown to contain framboids after having lain on the surface of a Mediterranean beach for as little as one to two years (Turner-Walker, 1993). The absence of bone degrading (collagen utilising) micro-organisms and the slow rate of hydrolysis in the Norwegian material is attributable to the low mean annual temperature of the burial environment and the fact that the bodies underwent much more rapid burial than is usual for organisms in the natural environment. The mean annual temperature in mid-Pleistocene East Anglia was certainly higher than that currently enjoyed in mid-Norway, which would imply that that the bones from the WRFWB examined to date experienced a relatively short exposure to aerobic micro-organisms and rapidly became enclosed in sediment that isolated them from both oxygen and extreme fluctuations in temperature. This must apply not only to the large, dense bones of the fossil elephant but also those smaller bone fragments found close to the top of the bed. This suggests that the body of water overlying the silts that now make up the Freshwater Bed, had low oxygen content, certainly at the sediment water interface. Of course it must be remembered that the huge overburden of the Cromer till and associated deposits that now make up the bulk of the cliffs at West Runton must be responsible for considerable compaction of the WRFWB, such that bones which now lie 10-20 cm below the top of the organic rich layer may originally have lain much deeper in the soft river mud.

CONCLUSIONS

Microscopic and X-ray diffraction studies of bones from the WRFWB confirm many of the mechanisms of pyrite diagenesis and subsequent pyrite oxidation products that have been reported by other researchers (Hannam *et al.*, 1996; Bottrell *et al.*, 1998). Backscattered SEM images of the early stages of framboid diagenesis indicate that pyrite grains form initially as discrete, well-nucleated spherical aggregates but progressive pyritisation results in the development of massive, crystalline pyrite. This pyrite can completely fill the available pore spaces in skeletal material and in some cases this has consolidated bones and ivory that has become fragmented, either by post-depositional movement of the surrounding sediment or other taphonomic processes such as trampling by large animals. Observations of the bones and sediment recovered during the 1995 elephant excavation suggests that pyrite in the sediment is more prone to oxidation than the pyrite within the structure of the skeletal material, possibly because the pyrite in the sediment is predominantly in the form of finely divided framboids, whereas the bone and ivory contains densely packed masses of framboids or massive pyrite. The XRD studies indicate that the oxidation of pyrite results in the formation of gypsum, from the reaction between acidic, sulphate containing solutions and hydroxyapatite, and melanterite, a hydrated iron sulphate ($FeSO_4.7H_2O$). Although the sediment in the upper part of the WRFWB, the oxidised Fe bed, is depleted in pyrite sulphur (Hannam *et al.*, 1996, Table 2) bone and ivory from this level retain both framboidal and massive pyrite within their pore structures, despite their exterior surfaces being obscured by rust coloured iron oxides. It is possible that in highly fossiliferous sediments, the pore structures of bones represent an effective sink for pyrite sulphur which is less susceptible to subsequent oxidation and leaching by groundwaters.

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NORFOLK SOILS AND SOIL MAPS

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ABSTRACT

The history of soil survey in Norfolk is reviewed and the changes in methodology described. Variation in soils is determined by geology and geomorphic processes and the different scales at which these factors operate is explained. It was from these bases that a soil sampling technique, to recognise and classify soils, and to delineate and define soil map units, evolved.

INTRODUCTION

Systematic soil surveys conducted by the Soil Survey of England and Wales started in Norfolk in 1959 and finished with the publication of the county map in 1982. Before this, in the 1930s, a small area in the fruit growing district around Wisbech had been mapped in some detail, at 1:25,000 scale. More importantly, the whole county had been mapped by Arthur Young at the very beginning of the 19th century. Young's work, published in 1804, is of great interest and still has value. His methodology consisted of recognising the main types of land use and relating these to soil texture. For the Wisbech map there is no record of the methodology used.

The earlier part of the more recent Soil Survey work was reviewed by Corbett (1971). At the very end of that review there is a brief mention of the mapping method and of some of the developments then in the pipeline. These developments led, in the later 1970s, to major changes in mapping technique. The object of this review is to put these on record in the context of an overview of Norfolk soils and their distribution.

When soil mapping started in Breckland in 1959, the Soil Survey had been in existence for thirteen years and there was some standardisation for the description and

Bull. geol. Soc. Norfolk (for 1998) **48**, 27-44 (Published 1999)

classification of soil profiles and of the mapping technique. Certainly in Eastern England the latter became a two stage process. The initial stage was sampling along transects across relief features to define and classify the range of soils and to give some insight into their distribution. In the second stage the sample points were the intersections of a regular 200 metre grid, and on this basis, map units were delineated. This procedure was used in Breckland and the area mapped, about 80 square miles, included all of the forest land and all SSSI sites. These land uses covered only parts of individual 1:25,000 sheets and the distribution of these partly mapped sheets along with the six and a half fully mapped sheets later surveyed are shown in Figure 1. The final 1:25,000 sheet to be mapped was TF82 (Helhoughton) and by this time, in the later 1970s, mapping techniques had been transformed.

Soil survey is concerned with two basic operations; (1) the recognition, definition and classification of particular types of soil called Soil Series; and (2) the plotting of their spatial distribution in soil map units. Traditionally, as on the 1:63,360 Cambridge Sheet (Hodge and Seale 1966), it is fair to say that, as the fieldwork progressed across the map sheet, both operations proceeded simultaneously. In practice, the problem was the sorting of data for both classification and map unit composition, when the data set was still incomplete; the full data set being available only on completion of the survey. The two stage process used in Breckland (as described below) was therefore an improvement.

The essence of the mapping procedure used on the TF82, Helhoughton sheet (Corbett 1987), was to divide the work into stages, with the data set for each stage being relevant to the whole of the map and to the achievement of a particular objective. In terms of overall progress on a map sheet, this means that when each stage is completed there is a certain level of understanding applicable to the whole sheet. Moreover, with succeeding stages, further levels of understanding are achieved but always for the map as a whole.

Before the improved Norfolk procedure can be explained in detail and its relevance judged, a broad outline of the soils, their different scales of spatial variation and the relationship of these to the nature and structure of map units, is needed.

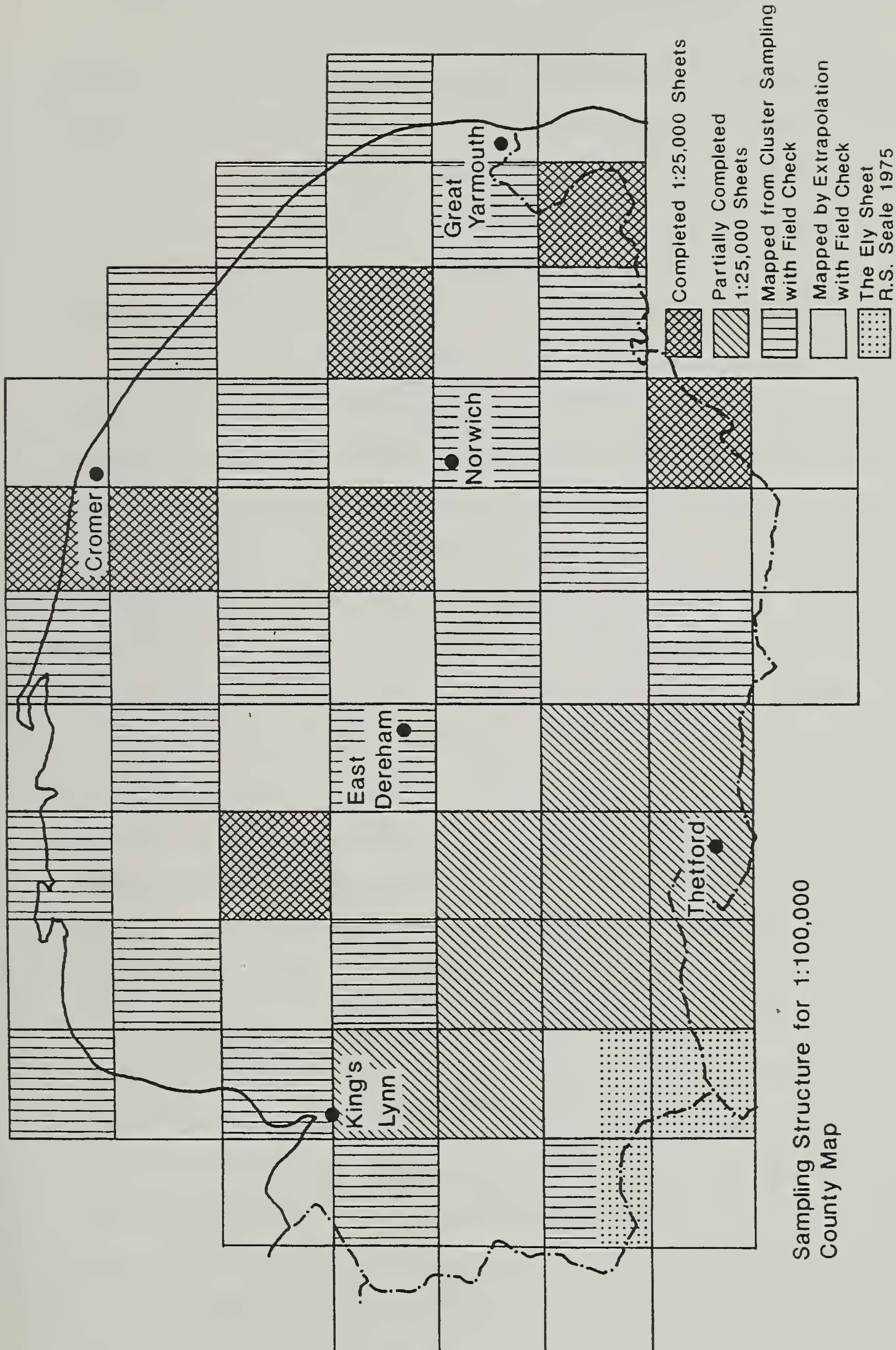


Fig. 1. Soil maps of Norfolk showing sample technique applied (see text for details).

THE SOILS OF NORFOLK: THEIR NATURE AND DISTRIBUTION

Soils and Geology

The soils of Norfolk and their distribution can be outlined in terms of their parent materials, i.e. the rocks or sediments from which they have been formed. A primary division is between the upland surfaces, mainly of Pleistocene age, and the Holocene (Flandrian) surfaces of Fenland and the coastal and estuary marshes.

The Pleistocene deposits are either basal tills laid down underneath an ice sheet, or glaciofluvial drift derived from its outwash. These deposits were laid down in the Anglian Stage of the Pleistocene, approximately four hundred thousand years ago (Rowe and Atkinson pers. comm.). There are two major till facies: (1) the chalky till (Chalky Boulderclay/ Lowestoft Till, CBC/LT) with chalk stones in a matrix mainly of Mesozoic clay and chalk flour (Perrin *et al.* 1979); and (2) the less calcareous North Sea Drifts with a matrix derived, at least in part, from the sandy deltaic deposits then flooring the North Sea (Corbett 1996). The CBC/LT forms a broad, flat, central plateau which rises to about 60 metres O.D., and creates a north-south watershed (Fig. 2, Unit 1.1). The NSD forms most of the lowland in the north east of Norfolk (Fig. 2, Unit 1.2). The northern parts of these lowlands are capped by glaciofluvial drift, the sands and gravels that form the surface over most of the feature known as the Cromer Ridge. The coastal platform to the north of the Cromer Ridge is cut into a range of materials including tills, NSD, Marly Drift (see below), and glaciofluvial outwash. From the western part of the Cromer Ridge, a gently graded outwash plain, the southern part, the Wensum Sands (Fig. 2, Unit 2.3), extends as far south as Norwich, separating the chalky till plateau to the west from the North Sea Drift lowland to the east (Cox and Nickless 1972). In addition, younger glaciofluvial sands and gravels form terraces flanking the Holocene floodplain in most of the major valleys. On the slopes above terraces, certainly on the eastern side of the central plateau, pre-glacial sand and gravel outcrop below the chalky till.

The CBC/LT with its predominantly clayey matrix is almost impermeable. However, on the north west side of the plateau, centred on the headwaters of the Wensum (Fig. 2, Unit 1.3), till with a higher content of chalk, particularly with flour or silt sized particles in the matrix - referred to in the literature as Marly Drift (Perrin *et al.* 1979) - is much more permeable.

Fig. 2. The soil landscapes of Norfolk simplified from Corbett and Dent (1993) showing generalised units mentioned in the text.

The generalised stratigraphic pattern described above applies to the uplands of central and east Norfolk but not to west Norfolk. On the north-west coast, from Wells to Hunstanton, loamy decalcified till (the Hunstanton Till), of Devensian age (10-60,000 B.P.), forms a low coastal platform (Straw 1960). To the south, between the western edge of the central plateau and Fenland (Fig. 2, Unit 2.2, the West Norfolk lowland), there is an escarpment of Cretaceous rocks: Chalk, Gault Clay and Lower Cretaceous sandstones, the last including both iron-rich cemented Carstone and the almost pure quartz sand Leziate beds. These Cretaceous beds dip gently to the east; however, further south, in south Norfolk, where their strike changes from NNW-SSE to NE-SW, their dip is almost horizontal. The Chalk here is capped on the high ground by thin chalky drift and forms, at 30 to 40 metres O.D., the low dissected plateau of Breckland (Corbett 1973). On the Breckland plateau the chalky drift is capped by up to a metre of aeolian sand, a coversand probably blown in by saltation from the area that today is the floor of the North Sea (Corbett 1996). Outside Breckland, to the east and north, on the central chalk till plateau, similar material is incorporated into the surface of the chalky tills. In north east Norfolk, on the North Sea Drift, the equivalent surface drift has a more silty loessal character (Catt *et al.* 1972; Perrin *et al.* 1974; Corbett 1977).

The Holocene surfaces of Fenland and of the coastal and estuary marshes have marked spatial variation in stratigraphy and lithology (Gallois 1979; Coles and Funnell 1981; Waller 1994). This stratification was caused by fluctuations in sea-level over the last 10,000 yrs. resulting in deposition of freshwater peats and marine mud, silts and sands (Coles and Funnell 1981).

In Fenland, the more southerly parts of the Romano-British silts and clays in the north, overlie at depth, thin layers of peat. These peats thicken southward and eventually crop out in south-west Norfolk. Over a large part of the outcrop, however, pump drainage, cultivation, weathering and wind erosion have destroyed the surface peat so that the underlying Fen Clay now forms the surface (Seale 1974).

On the east coast estuary marshes, the complete stratigraphic sequence is a five-layered sandwich with peat at the top, bottom and in the middle, separating beds of marine clays and silts (Coles and Funnell 1981). However, the surface peat on pump-drained sites under grass or arable cultivation has oxidised and the present land surface is the upper layer of marine sediment. In the downstream part of estuaries, this marine sediment spans,

or almost spans, the wide floodplains, but upstream, as it nears its landward limit, it tapers in towards the present river channel (Corbett and Tatler 1970; Hodge *et al.* 1984). In section, near the landward limit, the marine sediment feathers out over the underlying middle peat. This peripheral or feathering out zone marks the limit of pump drainage. On the undrained sites upstream, where the surface is still the upper peat, the vegetation includes alder carr, *Phragmites*, sedge and grass communities.

Marine sediments contain sulphides, and near the periphery of the drained zone, where the sediments are thin and decalcified, after drainage and oxidation, pH values well below 3.5 may occur (Dent 1986). Such values are particularly common on the floodplain of the River Ant. Where such low pH values occur beyond the spread of marine alluvium, as in the upper Waveney valley, the probable source of the sulphides is the Mesozoic clay in the matrix of chalky till (Catt pers. comm. in Corbett 1979).

Soils and Geomorphology

In the preceding section, variation in the soil parent material of Norfolk is shown to be related to the geology and spatial variation is measured in kilometres. However, further variation on the uplands occurs within these major divisions. There variations are described and defined in terms of the geomorphological processes which have affected surfaces since their deposition.

The relevant geomorphic processes operated at least during the Devensian, when the maximum ice advance just reached the north Norfolk coast. At that time, Norfolk experienced periglacial or tundra conditions, with frozen soil in the winter and surface melt in the summer. The nature of these processes is closely related to topographic position and the effect is spatial soil type variation on two scales; (1) a scale of tens to hundreds of metres; and (2) a scale measured in metres.

Solifluction operated at the larger scale causing summer downslope movement of melted surface layers over a still frozen subsoil (Corbett 1973). This occurred even on gentle slopes of less than two degrees and it has since been supplemented by present-day erosion. Cryoturbation operated at the small scale, i.e. the upward rupturing or mushrooming of subsurface layers into the weathered surface horizons following freezing and expansion. The centres of upthrust have a spatial interval of five to seven metres, and the effect of this process is most prominently expressed at the surface by soil and vegetation where the surface and subsurface layers have a contrasting lithology. The

Breckland landscape with its stripes and polygons is a good example of both solifluction and cryoturbation, and their interaction (Corbett 1973).

On the Breckland plateau, the aeolian or windblown coversand is up to 1 m thick, but on the peripheral slopes it is less than 30 cm thick. In section, and at depth, the stripes and polygons in the chalky drift are structurally similar (Fig. 3). The depths at which the structural similarity occur indicates that slopes have been truncated to a depth of 2 to 3 metres (Corbett 1973). This weathered or partially weathered soliflucted material has accumulated below the slopes on the floors of dry valleys. The full spatial range of the variation from plateau to valley floor caused by solifluction may extend over several hundred metres.

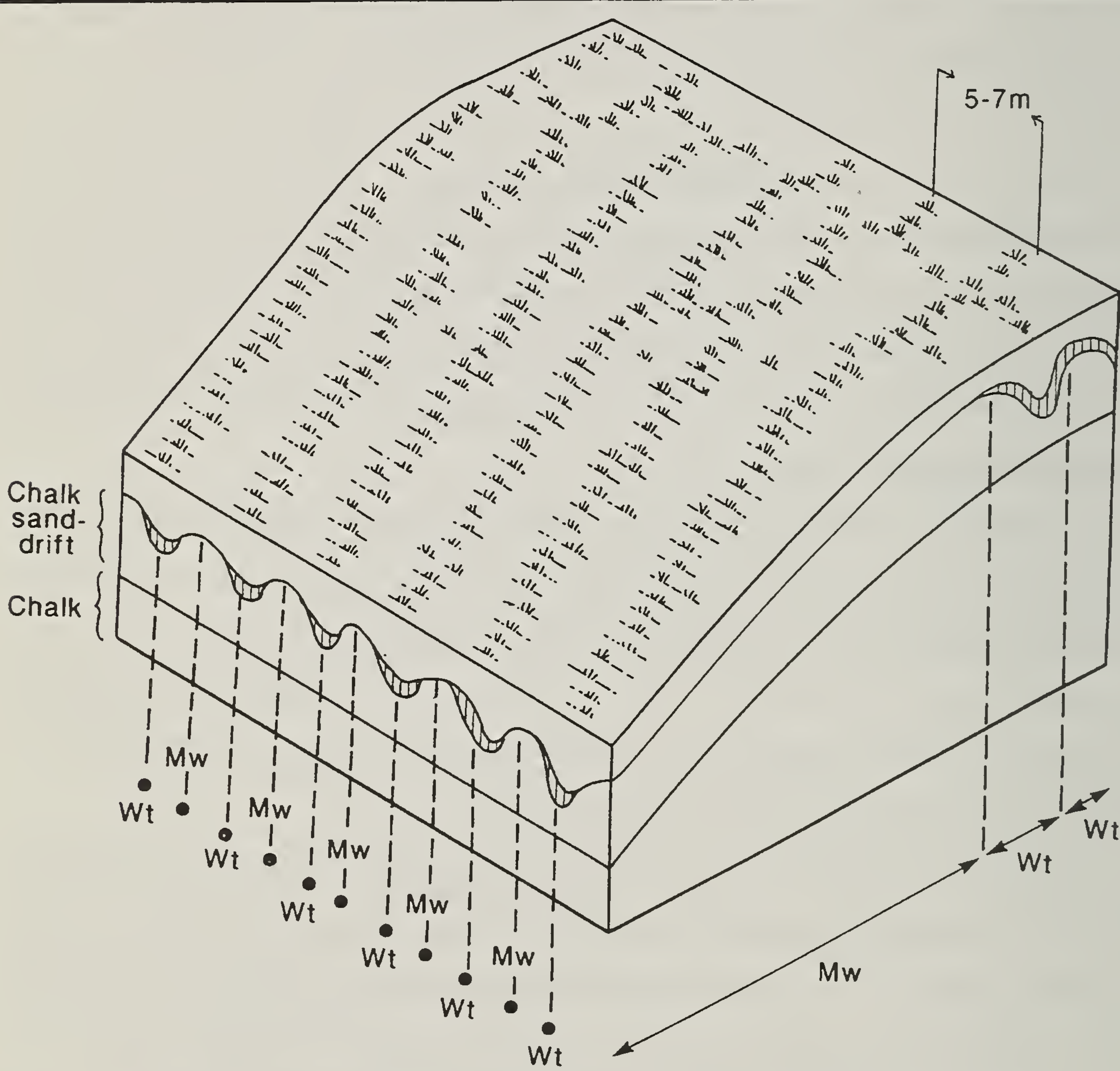


Fig. 3. Breckland stripes and polygons. Where Wt = Worlington Series (deep acid sand); Mw = Methwold Series (shallow calcareous sand). Note stripes on the slopes and polygons on the plateau. Distance between stripes and polygons is typically 5 to 7 metres.

The small-scale cryoturbation effect on the plateau has the form of isolated polygons, centres of chalk drift upthrust five to seven metres apart. These are separated by troughs of decalcified sand, more than 1 m deep. The equivalent features on slopes, where there has also been solifluction, have a stripe form. Subsurface ridges of chalky drift run down the slope, and are separated by roughly parallel troughs, again more than 1 m deep, of decalcified sand. Up until the arrival of myxamatosis in the 1950s, Breckland heaths were heavily grazed by rabbits and, under this grazing pressure, acidaphylous and calcicolous plant communities picked out accurately the distribution of the underlying periglacial features (Fig. 3). Since the disappearance of the rabbits these vegetation patterns have been swamped by plant communities with less specific site requirements.

On the chalky till plateau of central Norfolk, as in Breckland, periglacial features are represented by variation in the depth of coversand. However, in the surface layers of the Chalky till plateau, there has been much incorporation and intimate mixing of the coversand with the underlying clayey till. In general, the thicker or deeper the depth of incorporation of coversand, the coarser its texture. In the centre of the plateau the layer of incorporated coversand can be up to 1 m thick and its texture is coarse loam (Fig. 4). In contrast, on the very gently sloping land near the periphery of the plateau, coversand is represented only by a sand fraction in the plough layer, a fraction which is absent in the immediate subsurface clay. In spite of this sand, the plough layer texture is still clayey (Fig. 5). Between these two extremes most of the plateau surface has coversand incorporation to depths of 40-80 cm and the texture is fine loam (Corbett 1979).

The underlying chalky till with its clay matrix, is impermeable; in the winter months, when rainfall exceeds evapotranspiration, a perched watertable develops (Corbett 1979). However, on almost all sites, land drainage for arable farming holds this seasonal watertable at depths below 50 cm. At the edge of the plateau where the chalky till is <3 m thick and, throughout its depth, partially weathered and fissured, this seasonal watertable is only transient. On such peripheral sites the surface horizons have loamy texture and they represent the solifluction product from the chalk till plateau above. These surface loams, or head deposits, extend downslope from the weathered chalk till fringe and cap the upper parts of the preglacial sand outcrops on the sides of valleys. All of this spatial variation in depth and texture, from the centre of the plateau to valley sides, distances of several hundred metres, is interpreted as a solifluction effect (Corbett 1979).

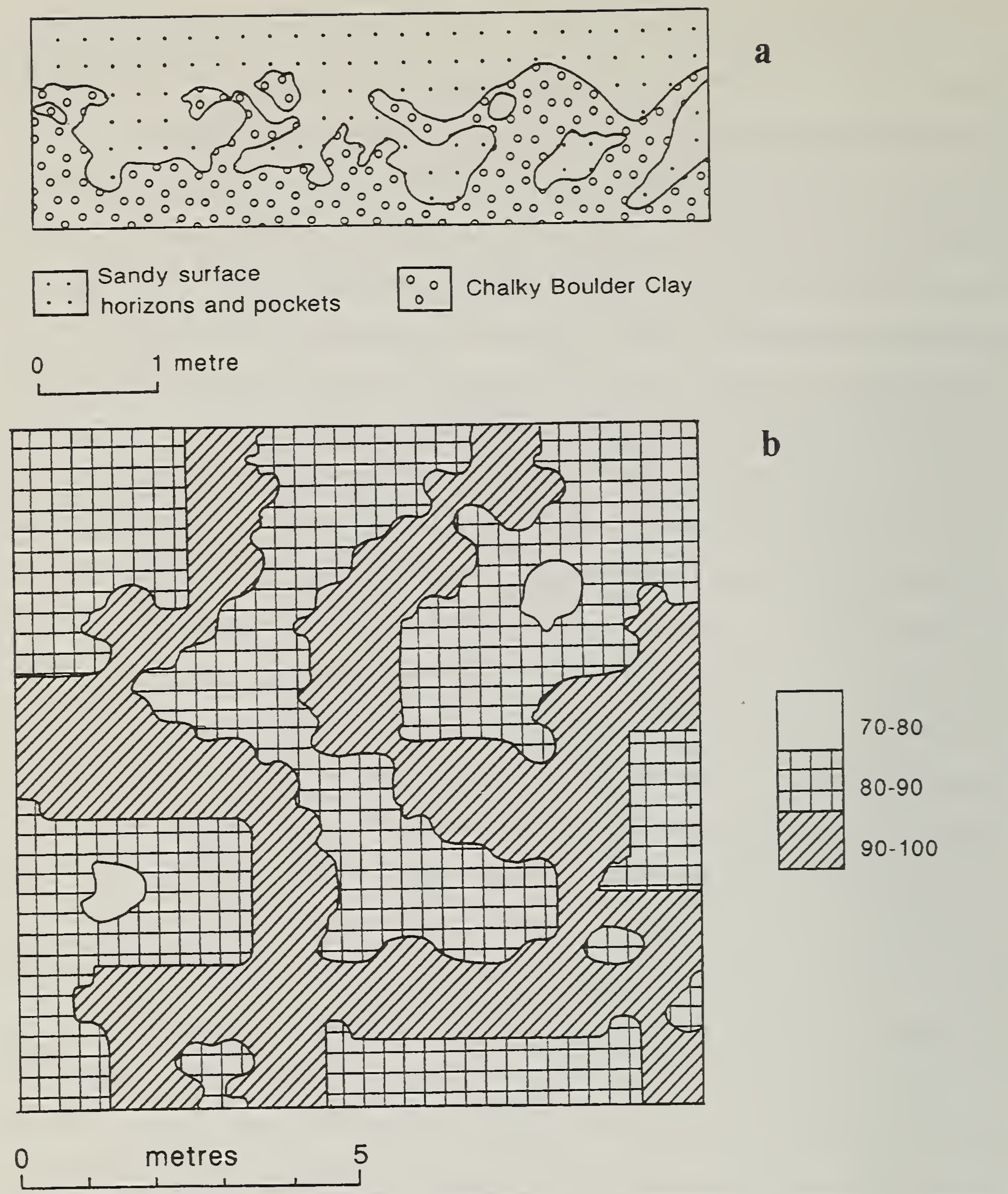
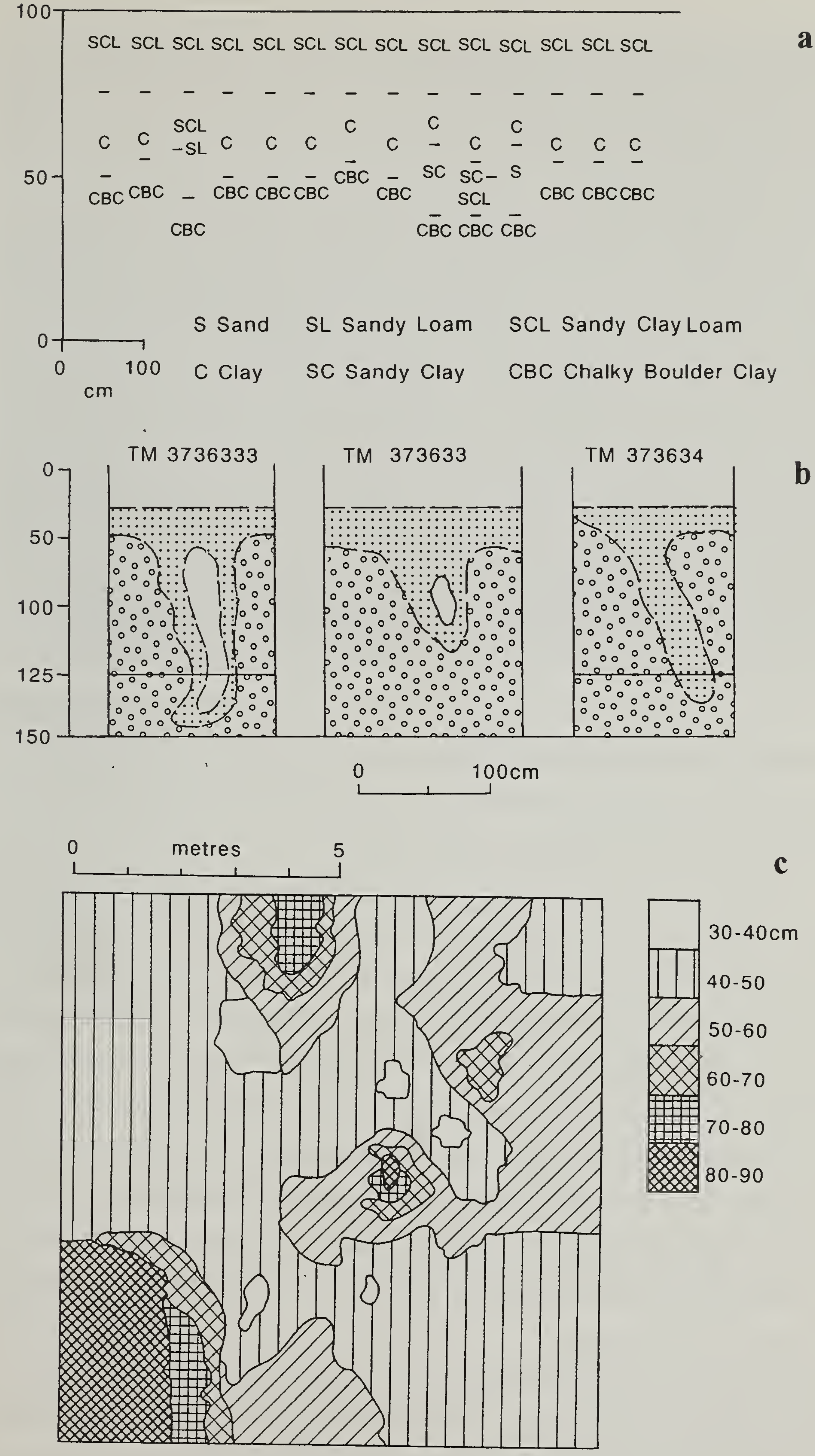


Fig. 4. (*Above*) Coversand patterns on CBC/LT at a central plateau site; (a) shows a cross section and (b) a plan view. Depth scale is in centimetres.

Fig. 5. (*Facing page*) Coversand patterns on CBC/LT plateau periphery site; (a) shows textural profiles in section based on hand augering; (b) shows sand pockets in section, and (c) shows plan views of sand pockets . Legend as in Fig. 4 (white area in centres of pockets are slightly differentiated sands).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Superimposed on the general trend described above, is variation in depth of coversand incorporation on a horizontal scale of a few metres. On central plateau sites with the deepest incorporation of surface coversand, even deeper troughs, filled with relatively pure coversand, interlock and isolate crests in the till surface; a coarse equivalent of the Breckland polygons (Fig. 4). Near the plateau periphery, with coversand represented only by the presence of a sand fraction in the plough layer, there are occasional shallow holes in the till surface immediately below the plough layer and these holes are filled with relatively pure coversand, (Fig. 5). Between the centre and the periphery of the plateau - i.e. over most of the plateau surface - the variation in depth again takes the form of more sandy troughs, but these do not interlock and, as far as the watertable is concerned, each sandy trough forms a small independent catchment a few metres long (Corbett 1979).

These local patterns on the chalky till plateau are not as symmetrical as those in Breckland. Moreover, between the three forms of pattern there is no sharp boundary, one type merging into the other, parts of a continuum. The subsurface features have little or no surface expression in landuse, site conditions being dominated by the hydrological condition which is common to the whole plateau. There is, however, some tendency for the occasional patches of moist oak woodland to be concentrated on central plateau sites, the most difficult sites for the drainage of surface water.

A third major landscape affected markedly by periglacial processes, is the glaciofluvial sands and gravels capped by silty coverloam in the north of north-east Norfolk; the Cromer Ridge. Here, solifluction and more recently, erosion, have stripped the loessal coverloam from most slopes. This material has accumulated in valley floors and re-entrants to depths locally exceeding 2 m. At such depths the coverloam is very silty, but near the surface, because of recent erosion, it becomes markedly more sandy. Above the slopes, on the gently rounded interfluves, the coverloam is 40-80 cm thick. The variation in depth occurs within a few metres and in plan, as shown on aerial photographs, it takes the form of a complex and intricate pattern of striations (Fig. 6; Corbett and Tatler 1974). These are interpreted as the combined effect of cryoturbation and solifluction, individual striations being local flow channels.

The reason why the striations are delineated so accurately on aerial photographs is the marked contrast in water-holding capacity between the underlying sands and gravels and the silty coverloam, the latter holding more than double the amount of plant available

Fig. 6. Differential growth in sugar beet reflecting variation in thickness of surface cover loam in northeast Norfolk, south side of the Cromer Ridge. (Photograph reproduced from Corbett and Tatler 1974, plate 1, courtesy of the Soil Survey. Crown copyright material is reproduced with permission of the Controller of Her Majesty's Stationery Office).

water. This site factor is reflected strikingly in plant growth, particularly in dry years when aerial photograph images clearly delineate, by darker tones, the more vigorous plant growth in valleys and re-entrants and on the interfluve striations (Fig. 6).

From these three examples of major soil landscapes in Norfolk it is clear that variation within them, caused by periglacial processes, is common to the whole of upland Norfolk. The surface expression of these variations, particularly with regard to plant growth, is most obvious where there is a marked contrast, chemically as in Breckland or physically as in north-east Norfolk, between the aeolian surface layers and the underlying drifts or tills.

Soil patterns on equivalent scales to those described above also occur on the Holocene surfaces of Fenland and on marshes. However, on these latter sites the process is purely depositional. During deposition, drainage channels, now fossil features termed

rodhams, had a sorting effect on the lithology of the alluvium. The land surface expression of rodhams is now low meandering ridges, and while the crests of these ridges, which enclose the original channel, tend to be silty and calcareous, towards the back swamp position they are less calcareous and more clayey (Seale 1975).

The ridges, ten to a hundred metres wide, now stand a few metres above the land surface. On Fenland peat they are particularly conspicuous and their distribution, as seen on aerial photographs, is a very fine branching network which coarsens towards the larger channels and rivers. On minerogenic marshes, tonal contrast on aerial photographs is slight but the rodhams can still be recognised in the field by their relief.

SOIL MAPPING TECHNIQUE USED ON 1:25,000 SHEET TF82 (HELHOUGHTON)

Taxonomic Sampling

Working from the basis that variation occurs on three quite different spatial scales, a three level nested sampling technique was devised. In this, 25 sampling centres were spaced in a regular two kilometre grid (Fig. 7), each sampling centre having four sub-centres 100 m north, south, east and west of its central point and each sampling sub-centre four sample points, 10 m north, south, east and west of its central point (Fig. 8). The sample site distances were derived from the spatial scales of variation described above.

The total number of sampling sites was four hundred (25 x 4 x 4) and at each the soil profile morphology was examined by augering, then recorded and classified. A taxonomic legend was then put together and each taxonomic unit plotted by symbol on the four 1:10,000 field sheets.

Soil Distribution and Map Unit Sampling

This stage used the traditional 200 m regular grid technique described earlier, but it was applied to only 36 of the 100 km^2 blocks on the 1:25,000 sheet. The 36 km^2 consisted of 9 blocks each of 4 km^2 (Fig. 7) regularly spaced so the blocks were 1 km apart. Within these blocks the 200 m grid sampling points were plotted on the base maps by symbol; the total number of sample points for the 9 blocks being approximately 1000.

On this basis, supplemented by the 400 clustered borings, soil patterns in relation to geology, geomorphology and topography could be interpreted and soil map units recognised, defined and delineated for each of the 9, 2 x 2 km blocks.

Soil Map Unit Extrapolation

On the same basis, geological, geomorphological and topographic, map unit boundaries could be interpolated or extrapolated between or from the 2 x 2 km blocks across the 1 km unmapped intervals between or around the blocks. This part was a desk exercise but delineated boundaries were then checked in the field. This rationalisation of the mapping technique reduced fieldwork time from 12 - 18 months to about 6 months and certainly improved the quality of the final map.

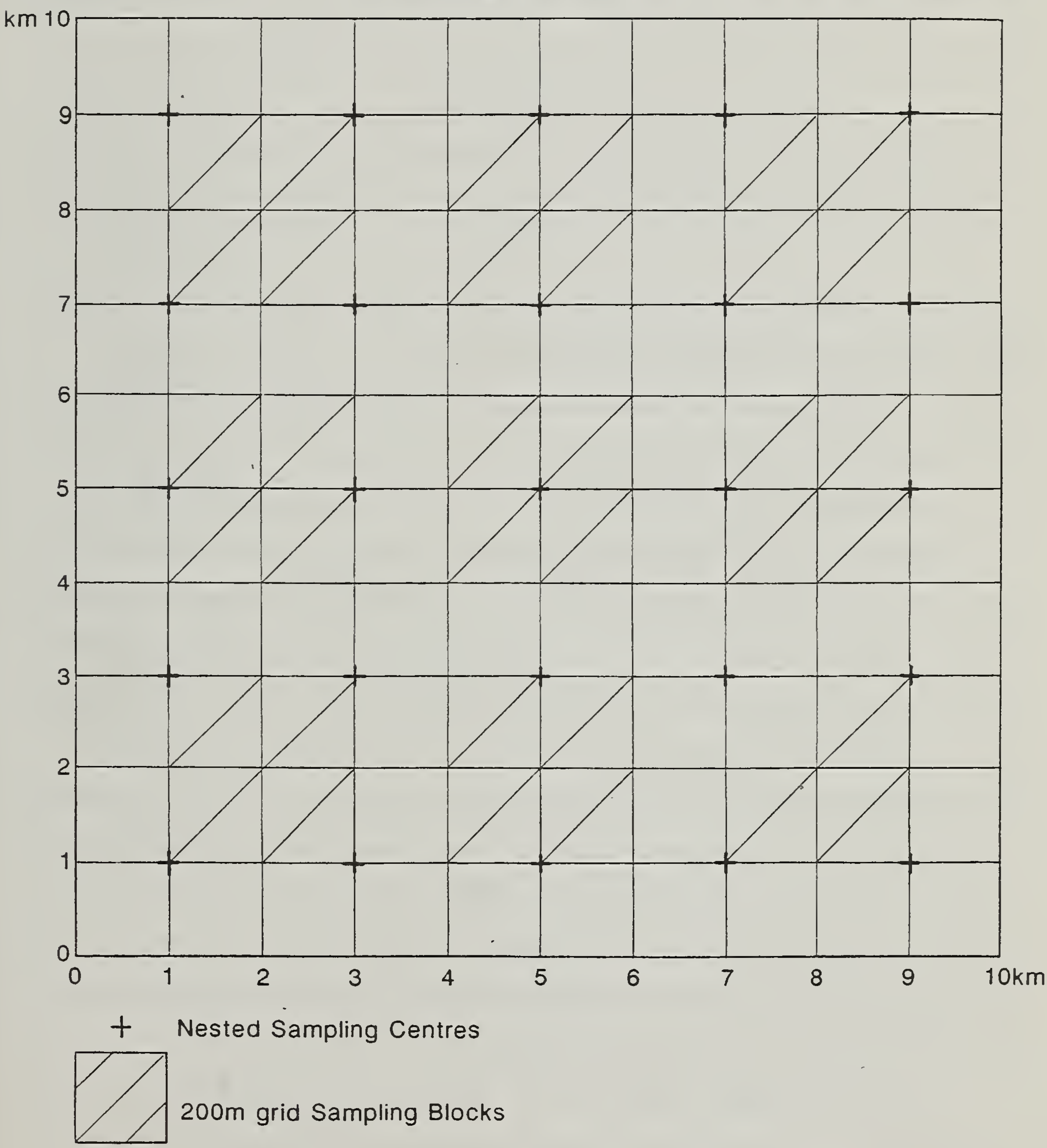

Fig. 7. Layout of nested sampling centres and blocks on the 1:25 000 sheet. Points of crosses represent the position of individual sample points as shown on figure 8.

Fig. 8. Layout of individual cluster sample points

STAGES IN THE PRODUCTION OF THE 1:100,000 SOIL MAP

The mapping technique for the 1:100,000 soil map was in many ways a continuation of that developed for TF82 (Helhoughton). Figure 1 shows the distribution of the completed 1:25,000 sheets for Norfolk and also those, including Breckland, where for various reasons, detailed mapping did not cover entire sheets. Further, in south-west Norfolk there is a small area which is part of the 1:63,000 Ely Sheet (Seale 1975). On the basis of this distribution of completed or partially completed detailed 1:25,000 sheets, alternating 1:25,000 sheets were selected throughout Norfolk (Fig. 1). Each of these alternating 1:25,000 sheets was then sampled by the 25 cluster technique used in the taxonomic stage of TF82 and on this basis alone, provisional soil map unit boundaries were delineated and later checked in the field. These soil map units were derived, with some amalgamations, from those on the completed detailed soil maps.

In the final stage, boundaries on the basis of geology, topography and geomorphology, were extrapolated from the detailed survey 1:25,000 sheets and the alternating cluster surveyed sheets into the intervening unmapped 1:25,000 sheets. These boundaries were then checked in the field to complete the 1:100,000 Norfolk Soil Map.

With regard to mapping technique, this was a one-off exercise. With some further simplification of map units this map was used again in 1984 as part of the 1:250,000 cover of England and Wales.

ACKNOWLEDGEMENTS

I wish to acknowledge the patience and accuracy of Rosie Cullington who word processed the typescript and Philip Judge and Sheila Davies who produced most of the diagrams.

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A POSSIBLE PRESENT-DAY PROCESS ANALOGUE FOR THE ORIGIN OF THE MARINE FAUNA OF THE LATE PLEISTOCENE MARCH GRAVELS OF THE FENLAND.

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ABSTRACT

Recent fluviatile sediments of Bathurst Island, Canadian Arctic Archipelago, contain a reworked marine fauna. The process producing the reworking may explain the origin of the marine fauna in the fluviatile March Gravels in Fenland.

INTRODUCTION

The March Gravels, a suite of Late Pleistocene (Devensian) shelly sands and gravels in Fenland, were first described in detail by Baden-Powell (1934). He interpreted them as marine gravels deposited during a submergence of the Fenland under a cold-temperate climate, post-dating an ice advance that deposited till in the Fenland area. He described from them a mixed fauna of marine and freshwater taxa.

Since Baden-Powell's original descriptions of numerous Fenland sites and of the sediments and their faunas, much further work on these gravels, their faunas and relation to the local Quaternary stratigraphy has been published. Gallois (1988) has given a full account of the March Gravels in the British Geological Survey Memoir for the Ely Sheet 173. He notes that the type section of the March Gravels is in the area of the March-Wimblington ridge (Fig. 1).

A number of problems arise when the stratigraphical relations and palaeontology of the March Gravels are considered. First, there is the problem of relating the gravels with marine shells to particular aggradations marked by terraces in the Fenland; secondly, there is the problem of how the gravels relate to the contemporary sea level; thirdly, there is the interpretation of the mixed faunas of the gravels, with marine and freshwater taxa.

Bull. geol. Soc. Norfolk (for 1998) **48**, 45-52 (Published 1999)

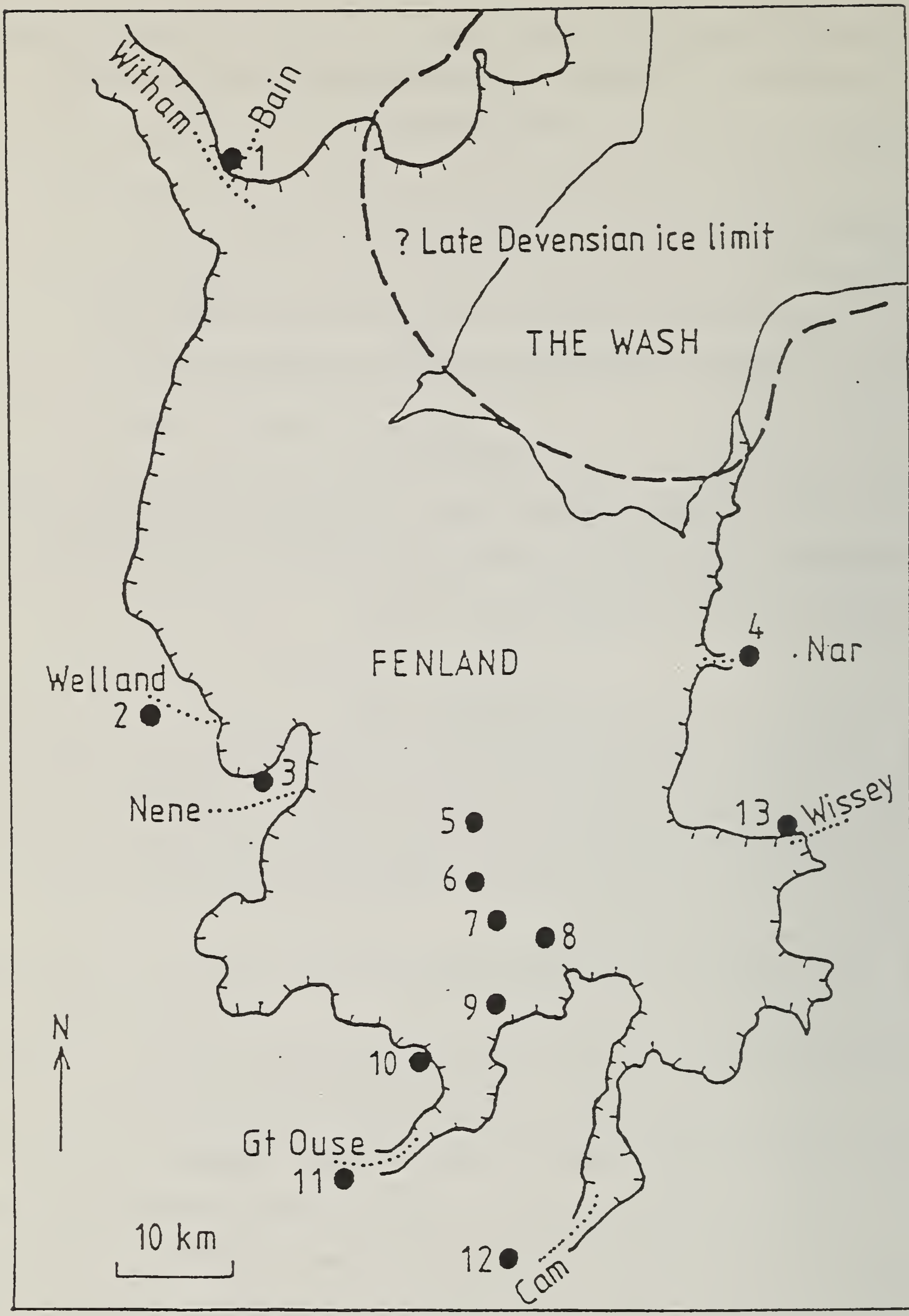

Fig. 1. Sketch map of Fenland showing position of sites mentioned in the text. Ticked line marks outer limit of Flandrian alluvium, peat and silt. 1 = Tattershall Castle; 2 = Maxey; 3 = Eye; 4 = Nar Valley; 5 = Graysmoor; 6 = March Town; 7 = Wimblington; 8 = Manea; 9 = Block Fen, Chatteris; 10 = Somersham; 11 = Galley Hill; 12 = Histon Road, Cambridge; 13 = Wretton.

These problems have been discussed by Keen *et al.* (1990) in respect of a site at Eye, Cambridgeshire, by Bridgland *et al.* (1991) in respect of the Peterborough district and by *West et al.* (1995a) in respect of sites at March, Wimblington and Chatteris, Cambridgeshire (see Fig. 1). The last paper suggested that the March Gravels were, in the (type) area considered, fluviatile sediments that overlay marine temperate sediments and contained a fauna with both marine and freshwater taxa, reworked from the older sediments during incision and aggradation by rivers draining the Fenland during the Devensian.

It is the general problem of the origin of the fauna of the March Gravels which is considered here. Were the faunas contemporary with gravel deposition or were they reworked?

COMPARING THE MARCH GRAVELS WITH SEDIMENTS FROM BATHURST ISLAND, CANADIAN ARCTIC

While studying plant communities of the polar desert of Bathurst Island, North West Territories, Canadian Arctic Archipelago, in the summer of 1995, I came across fluvial sediments with a marine fauna which may represent a present-day analogue for the deposition of the fauna of the March Gravels. Bathurst Island is split at 75° 45' N by a low-lying east-west valley, named Polar Bear Pass (Fig. 2). The valley has a superficial fill of Quaternary sediments, including silts and sands with a marine fauna (Blake 1964; Kerr 1974). Large lakes in the valley drain to east or west according to their position in the valley. The westerly lakes drain west into Bracebridge Inlet, with substantial streams cutting into the marine sediments (Fig. 2). These marine sediments are post-glacial and have emerged as a result of isostatic rebound in post-glacial times. They are low-lying members of a series of marine sediments, which indicate a marine limit of c. 90 m in the area of the east-central coast (Blake 1964).

Sections in the banks of westerly-flowing streams showed silts and sands with a marine fauna, including *Astarte borealis*, *Mya truncata* and *Saxicava arctica*, to a metre in thickness. These sediments slump into the flowing streams, where sorting takes place, with marine shells, entire and broken, being re-deposited along with gravel and sand in the stream bed within bar gravels or in the lee of sand ripples (Figs. 3-5). The result is fluvial sand and gravel containing a rich marine fauna.

The mixture of marine and freshwater faunas found in the Fenland March Gravels by Baden-Powell (1934) and in terrace gravels on the southern margin of Fenland at Somersham

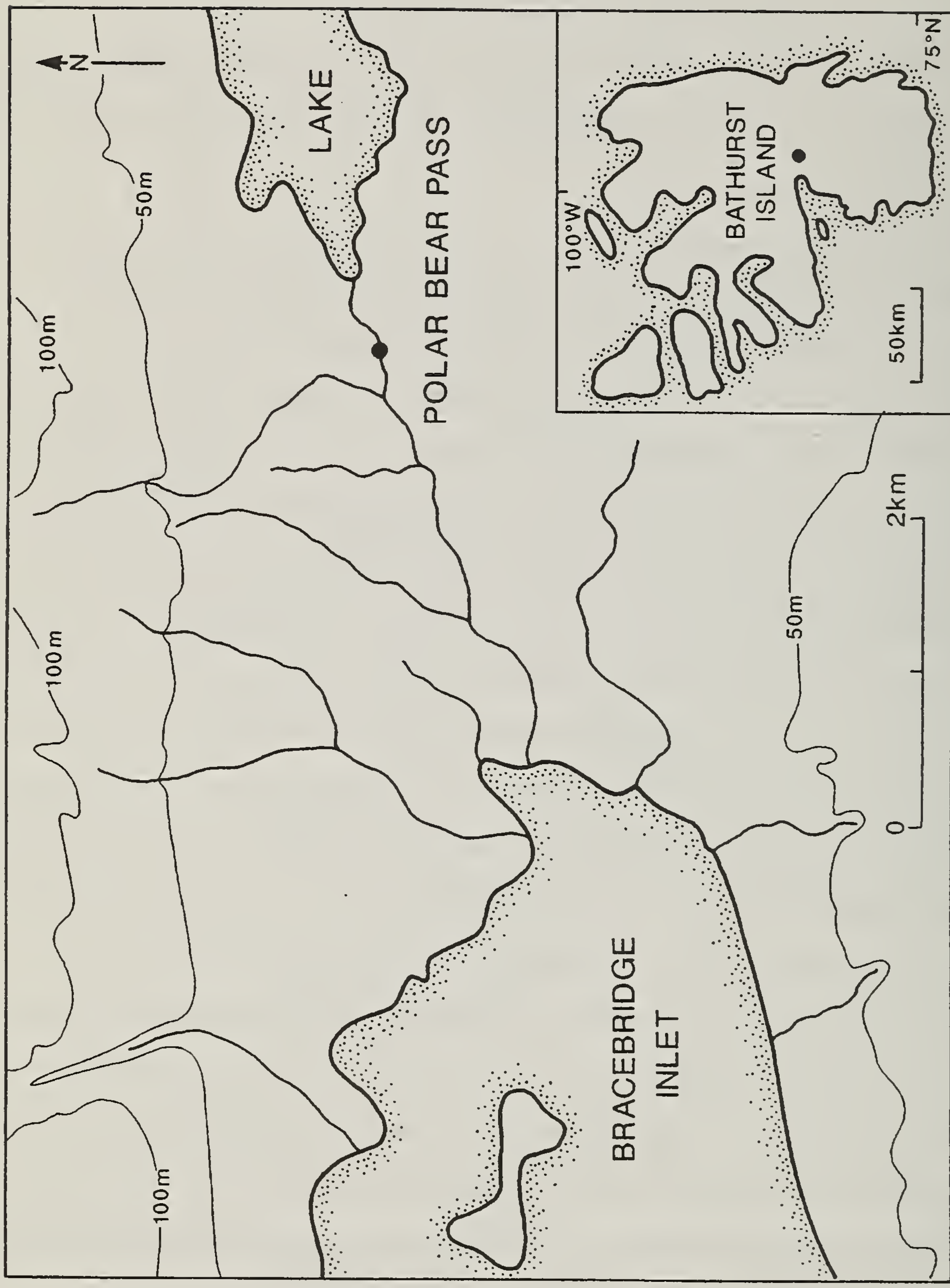

Fig. 2. A) Location of Bathurst Island and the study area. B) Western end of Polar Bear Pass. Areas with observed post-glacial marine sediments are hatched. Stream sampled at dot.

Fig. 3. Stream bank of marine sediments, Polar Bear Pass, Bathurst Island, North West Territories, Canada. A fauna of marine shells is reworked by erosion into fluvial sands and gravels of the stream bed in the foreground.

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by West *et al.* (1994) may be explained by a parallel process. The gravels form extensive terraces at heights from 5.5 m OD in the March area to 0 m OD at Block Fen. They are ascribed to the cold Devensian Stage, since they post-date temperate stage sediments at March, Chatteris, Wimblington (West *et al.*, 1995a), Block Fen (West *et al.*, 1995b) and Somersham (West *et al.*, 1994; Fig. 1). At Block Fen and Somersham they contain cold stage floras, with radiocarbon dates indicating a Devensian age. The underlying temperate stage sediments show highly fossiliferous freshwater and marine facies, the latter succeeding the former in a temperate stage assigned at present to the Ipswichian. Aggradation of gravel to 5.5 m OD at Chatteris follows freshwater temperate stage sediments. At March, Wimblington, Block Fen and Somersham (Fig. 1) aggradations to lower OD levels overlie marine temperate stage sediments.

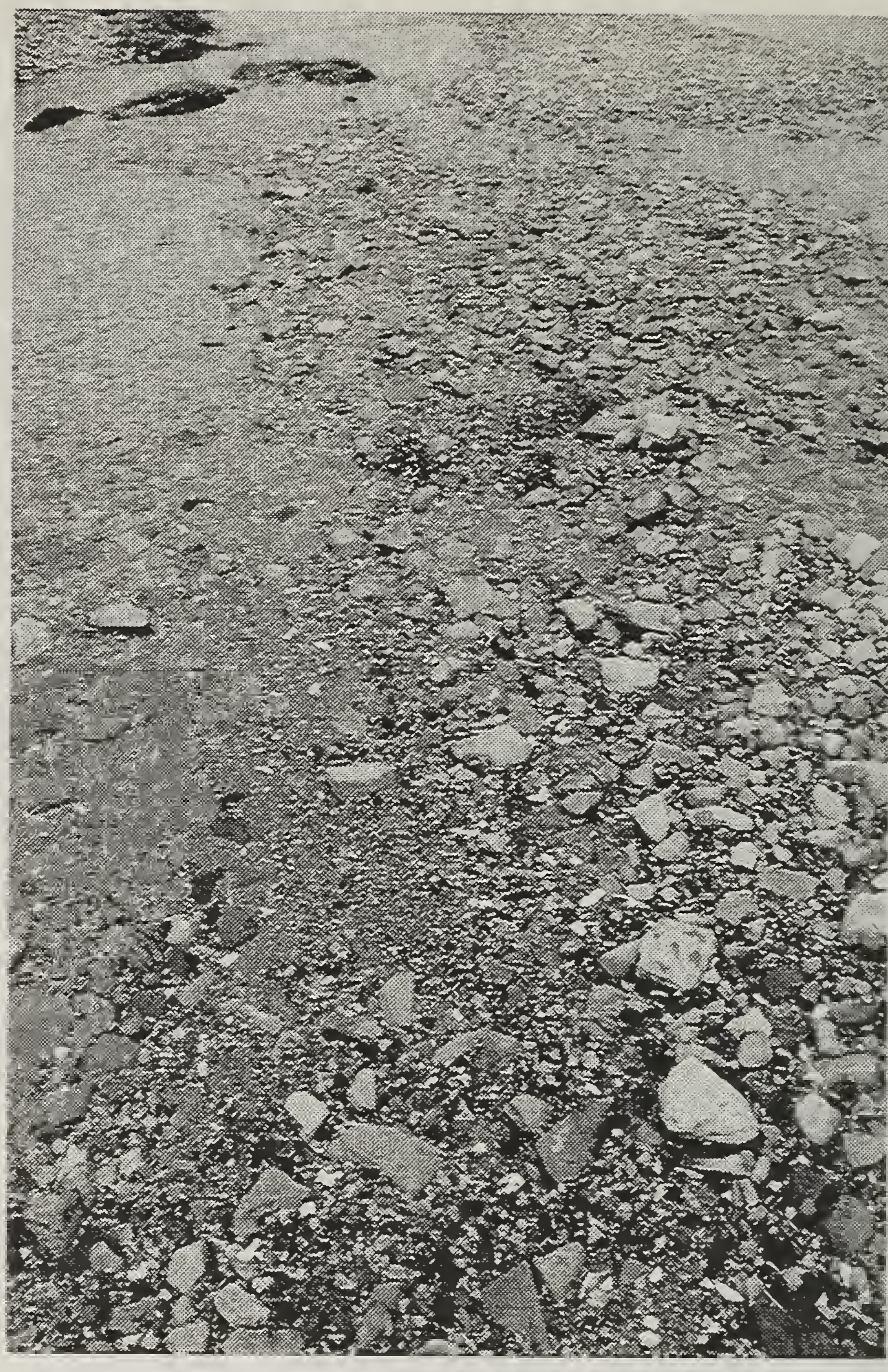

Fig. 4. Bar gravels with a marine fauna, adjacent to Figure 3 location.

The following sequence of events during the Devensian can be envisaged, leading to the incorporation of marine and freshwater faunas into the gravels. Aggradation of the higher level of gravels at Chatteris was followed by incision as fluvial erosion and deposition responded to falling sea levels of the Devensian. Incision resulted in the erosion of the temperate stage freshwater and marine sediments in the Fenland basin, with redeposition of the fauna in the gravels at times of aggradation, thus giving the mixture of faunas recorded by Baden-Powell and others.

Fig. 5. Fluvial sands with a marine fauna concentrated in the lee of ripples, adjacent to Figure 3 location.

CONCLUSION

The Bathurst Island sections illustrate the possibility and results of reworking when relative land-sea levels change. Similar effects may have been active in the Fenland basin as relative land-sea levels changed following the initiation of the Devensian Stage. The conclusion is that the taphonomy of fossil assemblages in fluviatile sediments must be assessed very carefully in relation to the regional stratigraphy before they are interpreted as being *in situ* or reworked, especially where there are problematical mixtures of faunas, as in the March Gravels.

ACKNOWLEDGEMENT

I thank Professor J. J. Donner for identification of molluscs.

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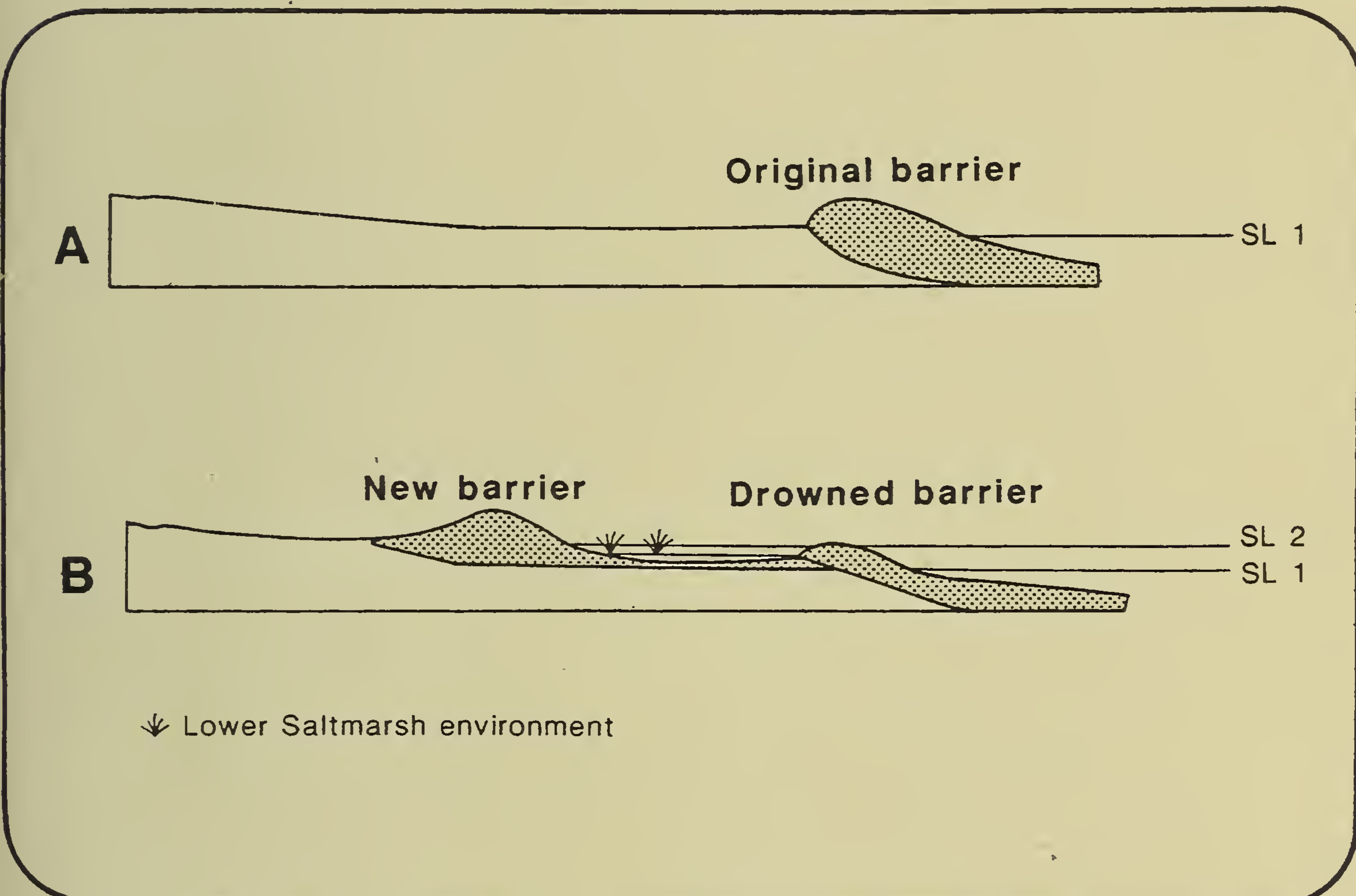
The illustration on the front cover is figure 1 from the article by Corbett in this issue of the Bulletin. It shows sampling structure used for the 1:100,000 county map (Norfolk).

ISSN0143-9286

BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

(FOR ARTICLES ON THE GEOLOGY OF EAST ANGLIA)

NO.49 1999



PUBLISHED 1999

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Trimingham, Norfolk

BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

No. 49 (for 1999) Published 1999

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EDITORIAL

Bulletin No. 49 concentrates on two aspects of coastal zone geology in Norfolk. The paper by Boomer and Woodcock suggests a model for the origin of the Stiffkey Meals, an enigmatic sand and gravel ridge found on the saltmarshes at Stiffkey in North Norfolk. Changes in sea level and possibly in climate are again implicated as possible driving mechanisms for the evolution of this part of our coast. The paper by Briant and others concentrates on the detailed sedimentology and provenance of much older coastal zone sediments exposed on the cliffed Norfolk coastline near Trimingham.

The appearance of Bulletin 49 brings the publication schedule completely up to date and I look forward to the submission of papers that will take to Bulletin into the new millennium. As usual I welcome the continued submission of papers on all aspects of East Anglian geology.

INSTRUCTIONS TO AUTHORS

If possible, contributors should submit manuscripts as word-processor print out accompanied by a disk copy. We can handle most word-processing formats although PC Word, WordPerfect or ASCII files are preferred. In addition we accept typewritten copy and will consider legible handwritten material.

It is important that the style of the paper, in terms of overall format, capitalisation, punctuation, etc. conforms as strictly as possible to that used in Vol. 41 of the Bulletin. Titles and first order headings should be capitalised, centred and in bold print. Second order headings should be centred, bold and lower case. Text should be 1½ line spaced. All measurements should be given in metric units.

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The editors welcome original research papers, notes or comments, and review articles relevant to the geology of **East Anglia** as a whole, and do not restrict consideration to articles covering Norfolk alone. All papers are independently refereed by at least one reviewer.

THE NATURE AND ORIGIN OF THE STIFFKEY MEALS, NORTH NORFOLK COAST

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ABSTRACT

Along the coast of North Norfolk there are a number of approximately coast-parallel sand and gravel ridges (or Meals) most of which protect the saltmarshes behind them from direct tidal inundation. They are particularly well developed at Stiffkey but are also known at Morston and Thornham. At Stiffkey Marsh they mark the boundary between the modern upper and lower saltmarshes. As part of a recent investigation into the Holocene (last 10,000 years) history of this coastline, a brief examination of the sediments which constitute the Stiffkey Meals and their relationship to the saltmarsh is made. The Meals only extend to about a metre below the modern marsh surface, indicating that they were formed relatively recently. They are composed of distinct layers of gravel, pebbles and sand, sediments which are generally found in much higher energy (more seaward) settings today. Immediately seaward of the Meals, fine-grained saltmarsh muds and silts to a depth of at least 5 metres occur, with a thin (0.5 metre) sand and gravel layer at less than 1 metre depth. This thin layer, records the transport of coarse material across the previous saltmarsh surface during what was a relatively short lived 'barrier emplacement' event approximately 1000 years ago. The mechanism behind the emplacement of the Meals may have been related to climatic deterioration or to sudden changes in the rate of relative sea-level rise along this coast.

Bull. geol. Soc. Norfolk (for 1999) **49**, 3-13 (Published 1999)

INTRODUCTION

This paper is one of a series which reports on a multidisciplinary investigation into the Holocene history of the North Norfolk coastline based on an extensive coring programme. The research was funded as part of the NERC community LOIS (Land-Ocean Interaction Study) programme. Under the LOEPS (Land-Ocean Evolution Perspective Study) component of LOIS a number of projects undertook investigations into the Holocene evolution of key coastal areas of Eastern England. North Norfolk was identified as an important area from the work of Pearson (1986) and Funnell & Pearson (1989). This paper also draws upon the unpublished work of Woodcock (1997).

THE STIFFKEY MEALS

The North Norfolk coast is commonly referred to as a barrier coastline and is dominated by two obvious geomorphological barriers, Blakeney Spit and Scolt Head Island. There are also aeolian dunes and a number of low-relief (<1.5 metres) coast-parallel barrier ridges, locally called 'Meals' or 'Meols' (e.g. Stiffkey Meals and Morston Meals, similar features are also seen further to the west at Thornham). These long and generally continuous ridges (Stiffkey Meals extend over three kilometres although they are transected by major channels) are undoubtedly natural in origin and should not be confused with the man-made seawalls at Holkham and Burnham. They often mark the boundary between upper and lower marsh environments since they restrict direct tidal inundation of the marshes. At Stiffkey they cause the upper marsh to flood and drain via a circuitous route through a few major channels (e.g. Cabbage Creek and Garborough Creek; Fig. 1). The Meals, up to 10 metres wide and standing about a metre higher than the upper marsh surface, are only over-topped during storm surge conditions.

The Stiffkey Meals ridge is composed of sand, pebble and well-sorted, clast supported gravel layers, resting on upper saltmarsh sediments. No other sedimentary features or structures were observed within the ridges. Funnell & Pearson (1989, p.31) state that the ridge is dune sand; however, the matrix of the ridge at Stiffkey is composed of coarse angular sands with pebbles and well rounded cobbles. This is clearly a water borne, rather than aeolian deposit, although it is possible that aeolian processes have subsequently modified the form of this topographically positive feature. Elsewhere along the coast these ridges do act as nuclei for dune sand.

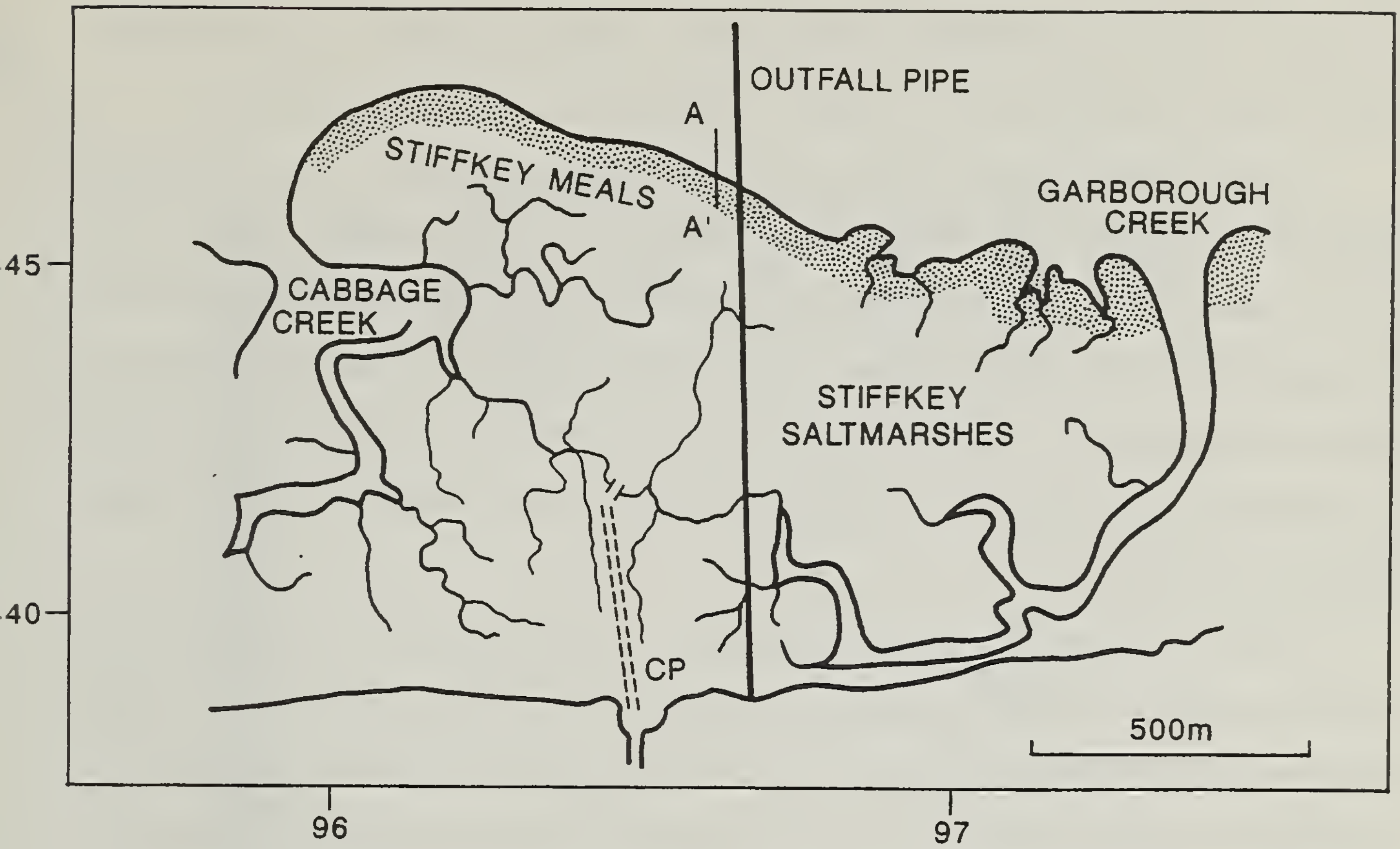

Fig. 1. Location map showing the Stiffkey Marshes and the transect (A-A') illustrated in figures 2 and 3. Stipple indicates the Stiffkey Meals ridge; CP = Car Park; Double dashed line = Track. Ordnance survey grid marks are shown for reference. (See also location on figure 5).

Ridges of this type have informally been referred to as 'cheniers', however, cheniers are generally found in sub-parallel series (the Norfolk Meals occur in isolation) on extensive mudflats associated with major estuaries and episodic sediment supply (Reading, 1986). Furthermore, many cheniers are comprised of biogenic shelly (molluscan) material unlike the Stiffkey barrier which is entirely minerogenic. Otvos & Price (1979) and Augustinus (1989) define cheniers as beach ridges resting on silty or clayey deposits which become isolated from the shore by a band of tidal mudflats, significantly, they are generally considered to form in progradational environments (i.e. where the shoreline is moving further out to sea). Otvos & Price (1979) and Augustinus (1989) also state that cheniers are mainly composed of sand grade material with some larger bioclastic components, whereas the Norfolk ridges contain significant amounts of gravel. Cheniers must represent relatively short-lived, extreme events (e.g. storms) which sort and deposit the coarse material above the normal tidal range. Recent studies indicate that for much of the past 8000 years the North Norfolk coast has been a predominantly

retrogradational system (shoreline moving landwards). Furthermore, Rhodes (1982) refers to cheniers being between 50-500 metres wide, clearly much larger than the Stiffkey Meals. Thus, we feel that the term 'chenier' is inappropriate in the present context.

At the present time, upper and mid-saltmarsh sediments are accumulating landward of the ridge, and lower saltmarsh sediments are accumulating seaward of it. Indeed, evidence from aerial photographs over the past 50 years suggests that the development of the lower marsh environment is spatially and temporally variable at Stiffkey, severe storm events can destroy years of accretion. During the early 1970s the marsh extended seawards but was largely destroyed over one winter in the early 1980s (Alan Gray, pers. comm.).

Recent coring behind, on and in front of the ridge (Figs. 2 and 3) indicates that the base of the ridge rests on upper saltmarsh sediments at about +1.50m OD (about a metre below the modern saltmarsh surface). There is a small lens of gravel at the same level in the saltmarsh sequence landward of the ridge, and a discontinuous layer of sand and gravel, mostly at about +1.50 to +2.00m OD, seaward of it, the base of which is interpreted as a transgressive wave ravinement surface (Nummedal and Swift, 1987), in effect the emplacement surface. Coring at the seaward margin of the present lower saltmarsh shows that saltmarsh sediments are present to a depth of at least 5 metres (Fig 2; core 96-17). The ridge is clearly resting on a thick saltmarsh sequence.

Micropalaeontological evidence from core 96-17 (Boomer, 1998) indicates that the sediments below the emplacement surface were deposited in an upper saltmarsh environment, flooded only on spring tides characterised by numerous saltpans and plants such as *Halimione*, *Limonium* and *Aster*, whereas those accumulating above it (i.e. the most recent sediments to accumulate) were deposited in a lower saltmarsh setting, i.e. flooded on most tides and characterised by plants such as *Salicornia* and *Spartina*. This supports the suggestion that the ridge formerly existed in a more seaward position, creating a much more extensive back-barrier setting where upper saltmarsh sediments accumulated. The event which moved the ridge material landward has redefined the lateral extent of both the upper and lower saltmarsh environments.

It is likely that some post-depositional sinking of the ridge into the underlying saltmarsh deposits has occurred. The 2.00 to 3.00m thickness of sand and gravel represents a significant loading on the unconsolidated saltmarsh surface sediments. The amount of subsidence has not been quantified precisely (approximately 0.40m), but should

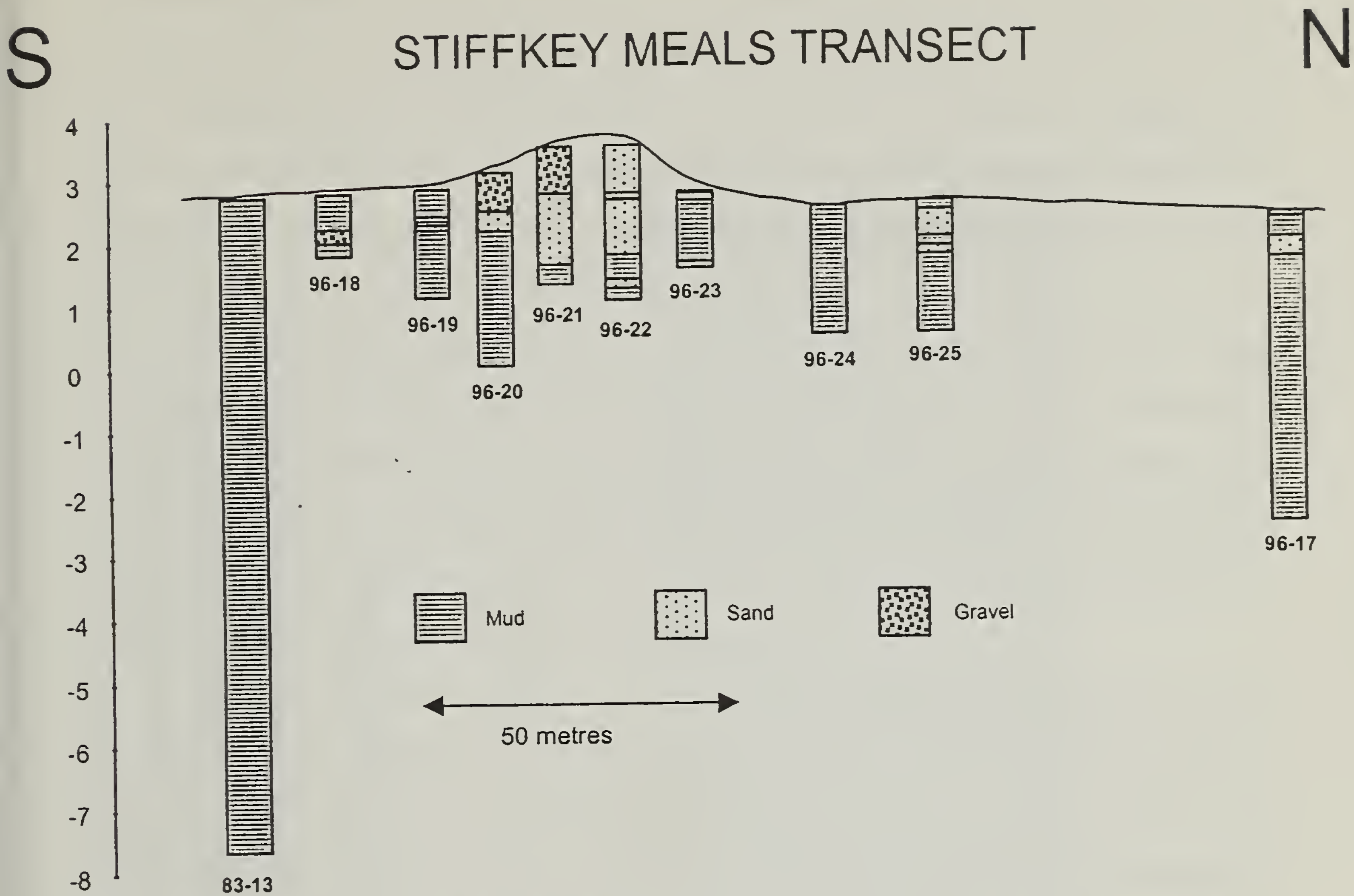

Fig. 2. Composite of 10 cores taken behind, through and in front of the barrier with the three main lithologies indicated. All cores were taken during the present research programme except for 83-13 which was recovered by Pearson (1986).

be borne in mind, when considering estimates for the age of emplacement of the ridge from it's base elevation. In this respect it is notable that the lower surface of the barrier is deepest under the thickest sequence of sand and gravel, where presumably the load is greatest.

The present-day level of the upper saltmarsh surface in this area (behind the Meals) is at about +2.70m OD. In order to calculate the original altitude of the saltmarsh surface at the time the barrier was emplaced we must look away from the main part of the barrier itself. Within the cores just behind the main ridge there are small layers of barrier material (Fig. 2, core 96-18) which formed as barrier material was washed onto the back saltmarsh. These 'washover' deposits are not sufficiently heavy to compact the underlying sediments and are, therefore, at about the same elevation now as when they were first deposited. If the level of the upper saltmarsh at the time of emplacement of the barrier is taken at the level of the oldest (i.e. lowest) washover deposits, we can estimate an initial

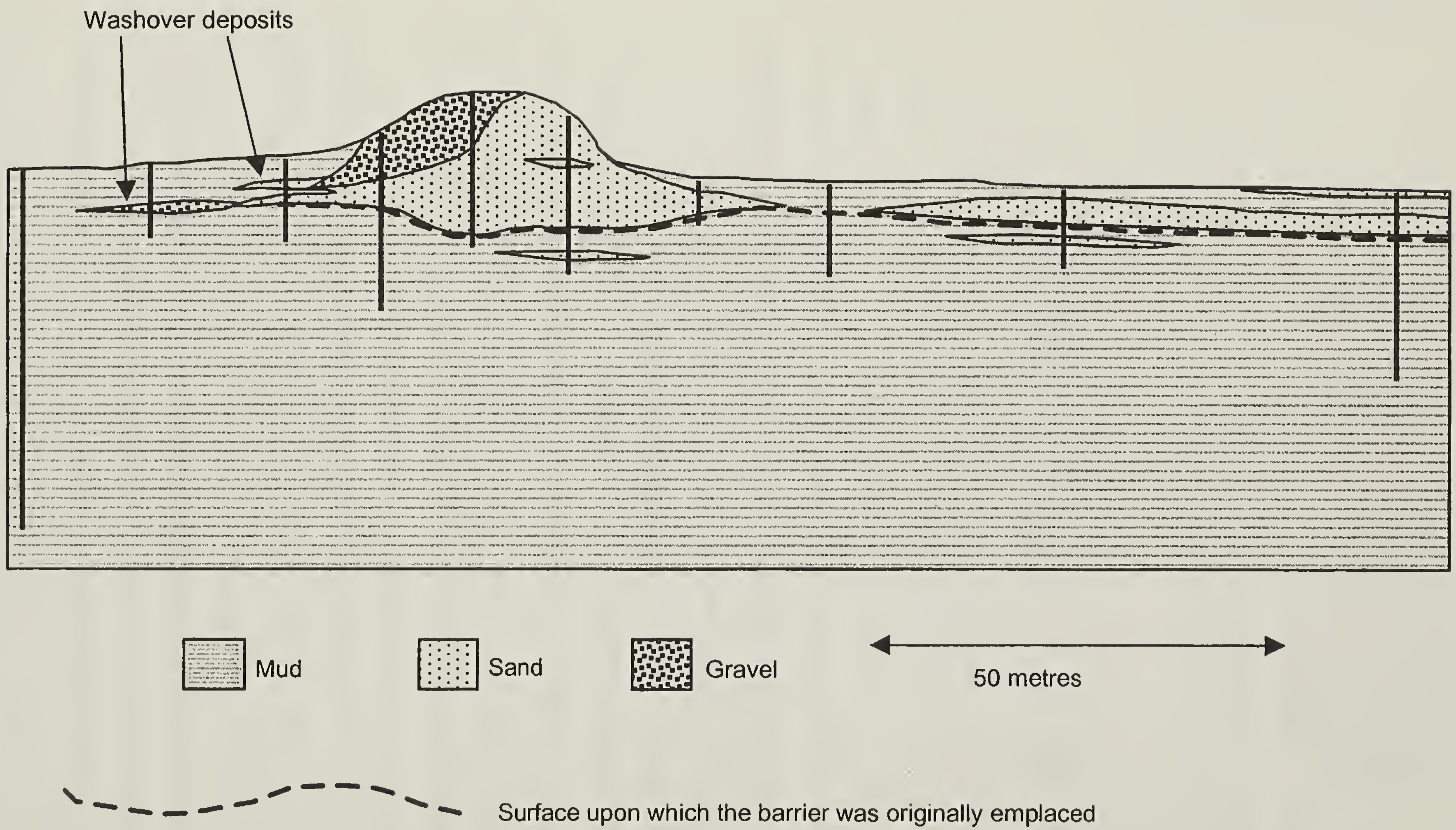

Fig. 3. Interpreted cross-section through the Stiffkey Meals transect with an indication of the surface upon which the original barrier was emplaced. Vertical black lines indicate the position of the cores detailed in figure 2. Gravels are clast supported and well sorted. The muds are largely minerogenic although many show evidence of rootlets.

saltmarsh elevation of about +1.60m OD. The difference between the modern saltmarsh surface (+2.70m OD) and the surface on which the barrier rests (+1.60m OD) is 1.10m (i.e. the saltmarsh has accreted vertically by 1.10m since the Meals first formed). The deepest part of the barrier (Fig. 2, core 96-21) lies at about +1.20m OD, if the barrier material was originally emplaced on a surface at about +1.60m OD then there has been maximum compaction of approximately 0.40 metres.

From the sediment-age estimation method published by Funnell & Boomer (1998) it is possible to estimate the date of barrier emplacement. The date is derived by comparing the elevation of microfossil assemblages in the sediment with the relative sea-level rise rate for this coastline. Funnell & Boomer (1998) determined the age of upper saltmarsh sediments at Morston, a few kilometres to the east, at a similar altitude of +1.70m OD as about 780 years BP (i.e. before 1950).

There is little direct evidence to suggest what conditions or events may have brought about the formation of this or other similar ridges along the Norfolk coastline, only the Stiffkey Meals have been investigated so far. We are not in a position to determine if their origin and emplacement in their present position resulted from the same event or if the barrier had previously formed in a more seaward position and subsequently moved (rolling-over) landward until reaching its present position. However, the model which best fits the available evidence is that of landward barrier migration following *in-situ* drowning (Fischer, 1961; Swift, 1975, Sanders & Kumar, 1975).

This model suggests that the Stiffkey Meals sediments accumulated initially as a feature more seaward (north) of their present position. They were subsequently transported landward either as a result of increased storminess or following a relatively rapid rise in relative sea-level drowning the offshore barrier (Fig. 4). This hypothesis is given weight by the observation today of an extensive linear sand (and probably gravel) ridge of low relief in front of the Stiffkey sand flats (about 2.5 km directly to the north) which may be a relic of a former barrier (Fig. 5). The mean low water mark (MLW) north of Stiffkey Marsh extends ESE towards Blakeney Point almost parallel with Blakeney Spit and it is possible that the dynamic forces which control the geometry of Blakeney Spit are also shaping the distal sand flats off Wells and Stiffkey. Aerial photographs of this area show clearly that the seaward margin of Bob Hall's Sand has a positive relief which suggests that it may be an incipient (or much degraded) longshore sand-spit almost in alignment with Blakeney Spit.

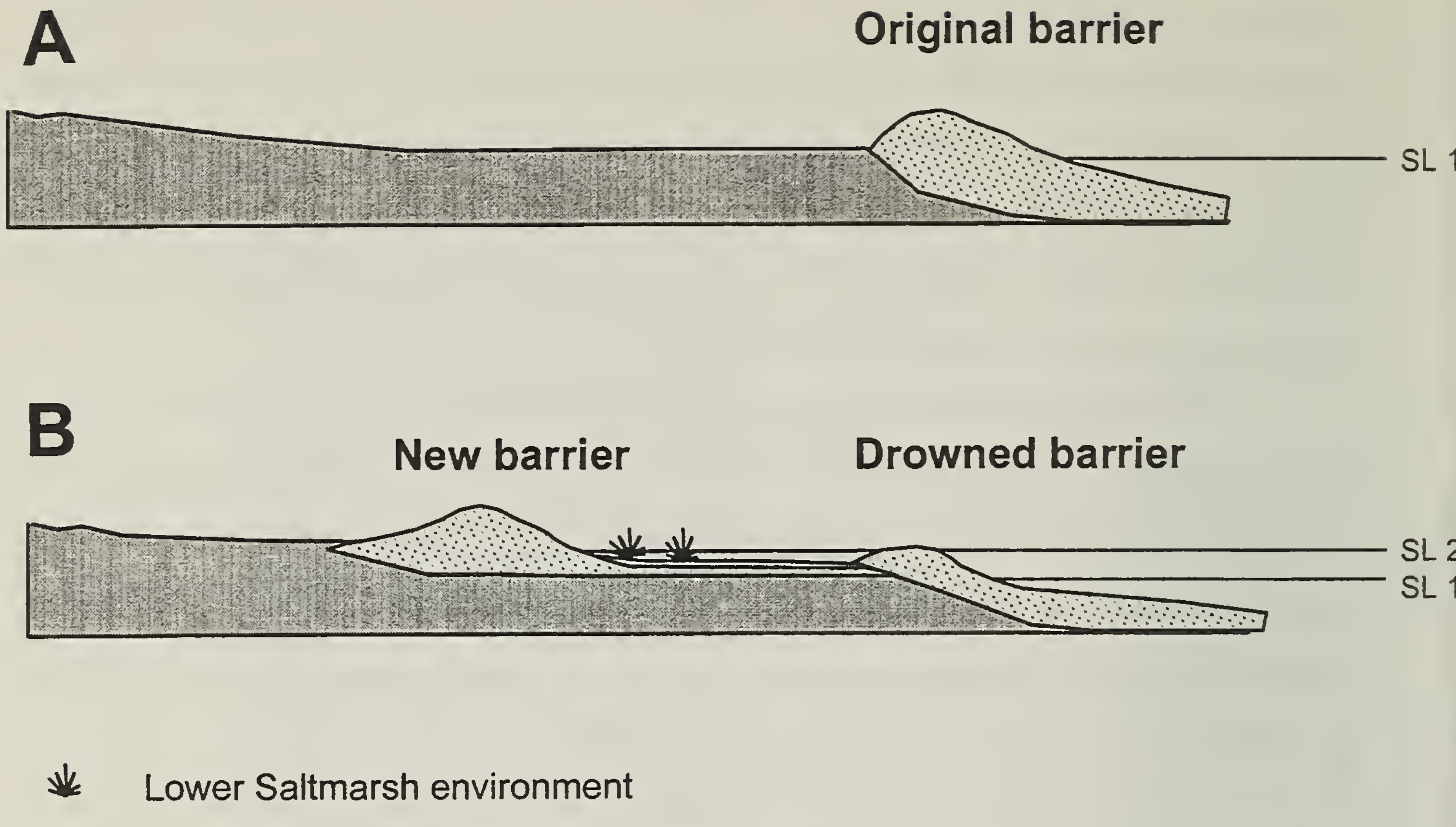

Fig. 4. Proposed model of barrier drowning and subsequent landward migration during enhanced relative sea-level rise (after Reading, 1986). **A** shows position of the original offshore barrier with the approximate position of the high water mark (HWM) at that time, **B** shows the position of the new barrier and new HWM.

Sanders & Kumar (1975) proposed that under conditions of accelerated relative sea-level rise, coastal barriers are initially drowned and subsequently migrate landward. The new barrier position is established and the former barrier may remain as a much diminished structure. Presumably, within this model the area between the old and new barriers would be strewn with a smear of clastic material sourced from the outer barrier, marking its passage across the marsh surface. This is much as we see today seaward of the inner barrier in cores 96-17 and 96-25 (Fig. 2). We suggest, therefore, that the inner barrier of Stiffkey Meals may have formed as a result of a short lived acceleration in sea-level rise at about 800 years BP. Whether all of the ridges along this coastline are of a similar age is unclear although, given their similar elevation this is possible.

The mechanism behind the formation of the Meals may be climatically controlled. From historical records we know that at about 800 years ago Britain was experiencing a relatively warm climate often referred to as the 'Climatic Optimum' which may have resulted in increased levels of sea-level rise. This was followed by a climatic deterioration. Our dating suggests that the Meals were emplaced towards the end of the Climatic

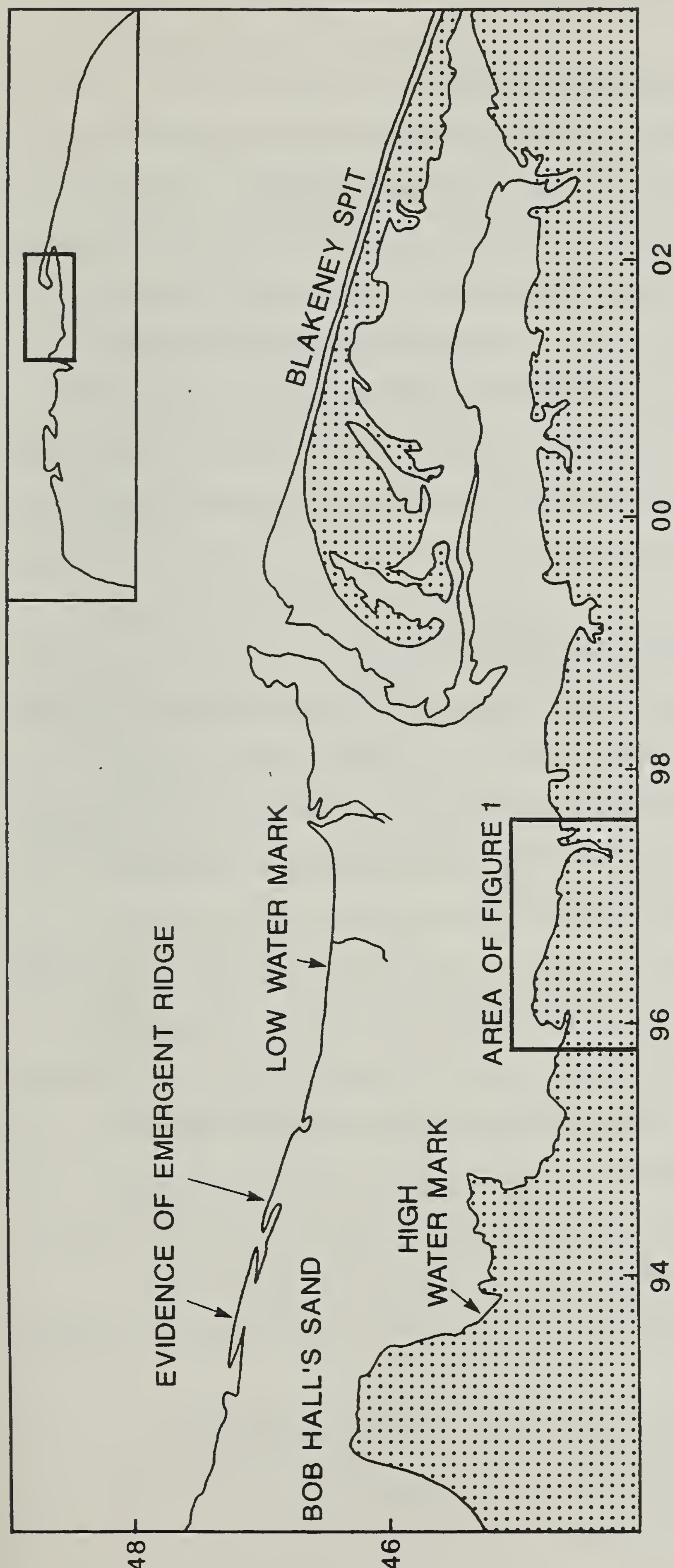

Fig. 5. Map showing location of Stiffkey Meals in relation to Blakeney Spit and degraded ridge visible at low water. Stippled area is above HWM. Ordnance survey grid marks are shown for reference.

Optimum (1100-1275AD) or early in the period of climatic deterioration and it is possible, therefore, that climate change may have played a role in their formation.

CONCLUSIONS

The Stiffkey Meals are composed of sands and gravels and rest directly upon saltmarsh sediments which extend to depth both in front of and behind the Meals. A thin veneer of sediments at shallow depth in front of the Meals record the migration of the barrier material across what was previously an upper saltmarsh environment. This 'event' has been dated to approximately 800 years ago. The Meals which we see today may be a degraded remnant of a once larger spit system which formed in higher energy conditions near the low water mark (currently about 2 km to the north). The material was probably transported to its present position following rapid drowning during a period of accelerated relative sea-level rise.

ACKNOWLEDGEMENTS

The authors wish to acknowledge the field assistance of Dr Greg Samways and Dr Rina Jones, we also wish to thank Dr Julian Andrews, Dr Barbara Maher and Prof. Brian Funnell (all UEA) for their support during the course of this work. Prof. Alan Gray and Dr Sue Brown (ITE Furzebrook) and Prof. Julian Orford (QUB) are also thanked for their comments. This is LOIS publication number 527.

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'PRE-GLACIAL' QUATERNARY SEDIMENTS FROM TRIMINGHAM, NORTH NORFOLK, ENGLAND.

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ABSTRACT

Pre-glacial sands and gravels are described from coastal exposures at Trimingham, near Cromer, north Norfolk and divided on the basis of sedimentary structures, particle size distribution, clast lithological content, pollen analysis and palaeocurrent directions into six sedimentary facies. The deposits are interpreted as having formed in a shallow marine environment by tidal current flow, with significant fluctuations in water depth and changes of current direction. Indicator lithologies show sediment sources to the coastal system both from the ancestral river Thames and from an as yet uncharted 'Northern' river. Pollen preserved in fine grained organic sediments in the lower part of the section, and ice-wedge casts from the upper part of the section suggest that the environment changed from a temperate-climate coastal zone fringed with alder-carr backswamp, to a permafrost environment with thermal-contraction patterned-ground. The temperate deposits formed during the Cromerian Stage and the permafrost structures as part of the early Anglian Barham Soil.

INTRODUCTION

Sands and gravels, silty clays and organic muds exist below the glaciogenic deposits of north Norfolk. They are known as Weybourne Crag or Cromer Forest-bed Formation, and attributed to the late Early, and early Middle Pleistocene (West, 1980; Zalasiewicz and Gibbard, 1988). The origin of these sands and gravels has long received attention (Reid, 1882, 1890; Harmer, 1896, 1909; Double, 1924) and the Weybourne Crag has traditionally been considered marine in origin, largely because of the presence of marine mollusca. However, West and Wilson (1966),

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West and Banham (1969) and West (1980) have demonstrated the occurrence of freshwater organic facies within the Cromer Forest-bed Formation, and Hey (1976, 1980, 1991), Hey and Brenchley (1977), Green *et al.* (1980) and Green and McGregor (1990) have suggested that 'pre-glacial' rivers flowed across the region. This issue has been the subject of recent debate (Hamblin and Moorlock, 1995; Green and McGregor, 1996; Rose, Allen *et al.*, 1996; Hamblin *et al.*, 1996), although recent research (Rose, Gulamali *et al.*, 1996; Rose, Lee *et al.*, in press) suggests that these rivers only contributed sediment to a coastal system in north-east Norfolk, but did not cross this region subaerially. The debate is compounded by the fact that exposures in the region are rare except along the coast. This paper contributes to this debate by providing a detailed description and analysis of sands and gravels beneath the glaciogenic deposits at Trimingham, east of Cromer, on the north Norfolk coast (Fig. 1).

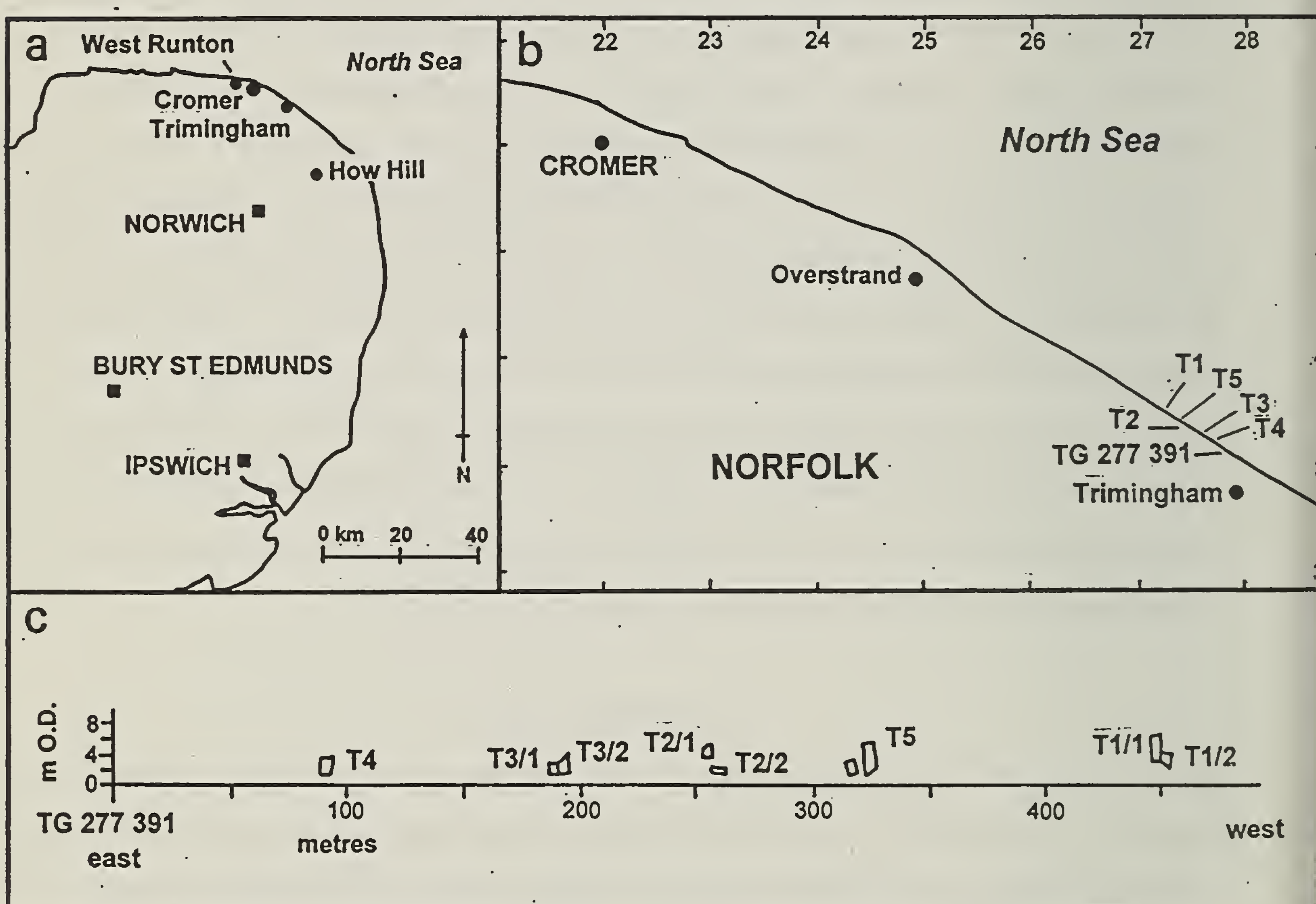

Fig. 1. (a) Location of Trimingham in East Anglia. (b) Location of sections at Trimingham, north Norfolk, England (TG 277391 - 274393). (c) Exact dimensions and location of sections.

SITE DESCRIPTION AND PREVIOUS WORK ABOUT THE SITE

The site at Trimingham, north Norfolk (TG 277391 - 274393) includes marine and freshwater sediments described by West (1980, pp. 56-58, fig. 29) overlain by 45-50 m of glaciogenic diamictons, laminated silts and clays and sands and gravels of the Anglian North Sea Drift Formation (Hart, 1988). The base of the sands and gravels extends below the present beach and West (1980) describes them as extending to 9 m below OD and records marine shells 7.5 m below OD. West (1980) interpreted the sediments as being marine and freshwater deposited in temperate and cool climates of the Pastonian and Cromerian stages of the Middle Pleistocene. Fish *et al.* (1998) recorded the presence of a periglacial composite wedge in the upper part of the sorted sands.

Sections are variably exposed depending upon coastal erosion and, more critically, the extent of land-slipping and toppling from the overlying glaciogenic deposits. West (1980) records that the section was poorly exposed during the period of his research and that for the period between World War 2 'until the 1960s' the area was a minefield. For this study, a total of six sections were studied and sampled along a 450 m stretch of coastline in December 1995 and summer 1998. Each section is approximately 4-6 m long and up to 5 m high. Altitude and position of sections studied during 1998 were surveyed with a Geographical Total Station (GTS) (Fig 1). The position of other sections is estimated.

METHODOLOGY

Each section was described precisely and sampled for particle size, clast lithology and pollen. Sediment description in the field consisted of detailed section drawing, moist Munsell colour determination, and measurement of depositional cross-sets to determine palaeocurrent direction. Particle size analysis and percentage organic carbon were carried out in the sediment analysis laboratories at Royal Holloway, University of London (RHUL). The >63 μm fractions were determined by sieving (Gale and Hoare, 1991), and the <63 μm fractions by the use of the Sedigraph method (Sedigraph 5100 and Mastertech, Coakley and Syvitski, 1991) (Table 1). Organic carbon determinations were carried out using the Walkley-Black procedure (Hesse, 1971). Clast lithological analyses were carried out on the 8-16 and 16-32 mm size fractions with the aid of a low-power stereo microscope. Results are expressed according to geological age which has relevance in terms of provenance (Rose, Gulamali *et al.*, 1996) (Table 2).

Table 1. Particle size composition of the 'pre-glacial' sands and gravels, Trimingham, north Norfolk. The units are arranged in stratigraphic order with the oldest at the base.

Sample No.	%Clay	(%Clay+ Silt)	%Silt	%Sand	%Gravel	Modes (phi size) (>clay)			Total sample weight (Kg)
Sediments infilling Wedge 1									
T3/2E		4.71		95.27	0.02		+4		0.11307
Sediments infilling Wedge 2									
T3/2J	12.93		17.35	69.54	0.19	+9	+4	-2	0.13987
Unit F									
T1/1E		4.61		93.26	2.13		+2	-3	0.12880
Unit E									
T1/1D		2.22		97.59	0.19		+3		0.15478
T1/1F		2.21		71.77	26.01		+1	-3	8.826
T1/2E		1.70		71.80	26.50		+1	-3	9.685
T3/2D		2.04		44.69	53.27		+2	-4	21.331
T3/2I		2.59		97.40	0.01		+2		0.10814
T3/2K		3.36		96.12	0.52		+1		0.13469
T4(1108)		0.20		99.80	0.00		+1	-1	0.19352
T5(1111)		0.10		97.85	2.05		+3		0.07307
Unit D									
T1/1B	7.04		2.87	90.09	0.00	+8	+3		0.10363
T1/1C	23.32		27.48	49.19	0.01	+8	+4		0.13654
T1/2C	5.39		3.19	91.39	0.02	+7	+4	-1	0.13350
T1/2D	9.50		22.38	68.11	0.00	+8	+4		0.12275
T2/1A	4.82		2.37	92.79	0.02	+8	+4		0.11359
T2/1B	15.98		16.17	67.47	0.38	+8	+4		0.10558
T3/1D		3.48		95.77	0.75		+3	-1	0.14398
T3/1E		3.35		96.57	0.08		+4		0.11266
T3/2F		3.33		96.65	0.01		+4		0.14704
T3/2G		0.93		99.06	0.01		+2		0.12370
T3/2H	5.01		8.10	86.54	0.35	+7	+3		0.15047
T3/2L	8.71		5.47	85.64	0.18	+8	+4		0.10394
T3/2M		0.99		99.01	0.00		+2		0.10728
Unit B									
T1/1A		0.69		99.19	0.12		+2		0.15094
T1/1H		2.32		91.50	6.19		+2	-3	0.18081
T1/2B		0.82		94.77	4.41		+2	-3	0.16604
T2/2A		1.17		98.83	0.00		+2		0.11154
T2/2B		1.15		98.85	0.00		+2		0.14087
T3/1B		1.05		96.76	2.19		+1	-2	0.16195
T3/1C	44.17		27.89	27.85	0.09	+9	+4		0.12533
T4(1107)	0.27		2.62	62.90	34.30	+6	+2	-4	7.345
T4(1109)		0.12		90.94	8.94		+1		0.12992
T5(1104)		1.13		84.66	13.87		+2	-3	8.912
T5(1103)		1.39		50.82	45.83		+2	-3	4.803
Unit A									
T1/2A	6.62		7.46	85.92	0.00	+8	+3		0.11333
T3/1A		1.57		98.43	0.00		+3		0.10255

Organic samples were prepared for pollen analysis according to the standard procedure used in the palynology laboratories at RHUL, which is an adaptation from Moore *et al.* (1991). This method involves sieving to retain the 5-90 μm fraction; heavy liquid separation with sodium polytungstate (specific gravity: 2.0 gm cm^{-3}); acetolysis and mounting using glycerol jelly. Hydrofluoric acid treatment was not carried out since little inorganic residue was present. Between 150 and 200 identifiable land pollen grains were counted for each sample and traverses covered the entire slide to ensure representativeness. Identifications are according to Moore *et al.* (1991) supported by the use of the RHUL Geography reference collection. Frequencies are calculated as percentage of identifiable land pollen per sample. Land taxa are expressed as a percentage of this sum, and aquatics, spores, and dinoflagellates as a percentage of this sum plus the total of the respective group, following the procedure used in this area by West (1980). Results are expressed in tabular format in taxonomic order according to Stace (1991) (Table 3).

SEDIMENT DESCRIPTIONS

Six facies units have been identified, with A at the base and F at the top.

Sections T 1/1 and T1/2, (Figs 2 & 3)

Unit A is exposed on the western side of section T1/2 and is <1 m thick. It is a light yellowish and olive brown to yellowish brown (2.5Y6/4, 2.5Y5/4, 10YR5/6) fine sand with a modal particle size of 125-250 μm (Table 1). Structures are picked out by light grey silt layers, which grade from in-drift climbing ripples through a massive structureless sand to in-phase climbing ripples. Above this, the beds are deformed and the unit is truncated at the top.

Unit B consists of ≤1 m of medium-coarse sands and fine gravels with modal particle sizes of 250-500 μm and 8-16 mm. The sands range in colour from pale yellow and light yellowish brown to pale olive (2.5Y7/3, 2.5Y6/4, 5Y6/3). The lower part of the unit consists of large-scale tangential cross-beds which include gravels or silt and clay lenses along the lower part of the bedding planes. These large-scale cross-sets are replaced by small tangential sets with an amplitude of 20-30 cm resting on multiple, concave erosion surfaces that truncate the lower beds. Well developed trough cross-bedding is seen at the top of the unit. Palaeocurrent measurements give readings that range from 324-112°, with a modal grouping from 324-48° (NNW-NNE). Dips are ≤28°. The top of the unit is truncated.

Table 2a. Lithological content of the gravels at Trimingham, north Norfolk, expressed as percentage of total sample number (n). Samples are arranged according to stratigraphic position with the oldest at the base.

Site No	n	Carb. chert	Triassic qzt	v.qz	schorl	total	Jurassic Rhaxella chert	sst. lst+ irnst+ shell	total	Cretaceous sst. lst+ irnst+ shell	glauc sst	Green-sand chert	chalk	flint	total	Tertiary/ Pleist. chatter-marked flint	Total Flint	Ign + Meta	Unk.
(8-16 mm size range)																			
UNIT E																			
T1/1F	1776	**1.9**	9.8	13.3	0.7	**23.8**	0.9	0.0	**0.9**	0.3	0.0	1.0	0.0	62.7	**64.0**	**9.2**	**72.0**	**0.0**	**0.2**
T3/2D	1448	**0.5**	15.5	18.9	1.9	**36.3**	0.5	0.0	**0.5**	0.0	0.0	0.3	0.0	50.9	**51.2**	**11.3**	**62.2**	**0.2§**	**0.1**
UNIT D																			
T2/1C	1081	**1.0**	9.7	22.0	0.6	**32.3**	0.5	0.0	**0.5**	0.0	0.0	0.2	0.0	55.3	**55.5**	**10.4**	**65.7**	**0.0**	**0.4**
UNIT B																			
T1/1H	278	**2.9**	5.8	14.4	1.1	**21.3**	1.8	0.7	**2.5**	1.8	0.0	1.1	0.0	65.8	**68.7**	**4.7**	**70.5**	**0.0**	**0.0**
T1/2G	454	**0.4**	13.4	20.9	0.4	**34.8**	0.2	0.0	**0.2**	0.0	0.0	0.9	0.0	52.4	**53.3**	**11.0**	**64.3**	**0.0**	**0.2**
T4/1	423	**3.1**	12.7	14.8	0.0	**27.6**	3.1	0.0	**3.1**	0.0	0.0	0.0	0.0	56.7	**56.7**	**7.8**	**64.5**	**0.0**	**1.7**
T5/1	1040	**7.3**	10.1	20.0	0.8	**30.9**	0.9	0.0	**0.9**	0.0	0.0	0.0	0.0	47.9	**47.9**	**11.4**	**59.3**	**0.0**	**1.6**
(16-32 mm size range)																			
UNIT E																			
T1/1F	104	**0.0**	10.6	9.6	1.0	**21.2**	0.0	0.0	**0.0**	1.0	0.0	1.0	0.0	51.0	**53.0**	**26.0**	**79.0**	**0.0**	**0.0**
T3/2D	297	**0.3**	19.2	9.4	0.0	**28.6**	1.3	0.0	**1.3**	0.0	0.0	0.0	0.0	45.1	**45.1**	**24.6**	**69.7**	**0.0**	**0.0**
UNIT D																			
T2/1C	63	**1.6**	22.2	9.5	1.6	**33.3**	0.0	0.0	**0.0**	0.0	0.0	2.0	0.0	42.5	**44.5**	**20.6**	**65.5**	**0.0**	**0.0**
UNIT B																			
T1/1H*	128	**2.3**	14.1	11.7	0.0	**27.3**	0.0	0.0	**0.0**	0.0	0.0	0.0	0.0	42.2	**42.2**	**28.1**	**70.3**	**0.0**	**0.0**
T4/1**	114	**0.7**	10.4	16.0	1.4	**27.8**	1.4	0.0	**1.4**	0.0	0.0	0.0	0.0	41.0	**41.0**	**27.1**	**68.1**	**0.7#**	**1.3**

* includes T1/2G
** includes T5/1
§ acid volcanic
\# basalt

Table 2b Distinctive lithological sub-fractions at Trimingham, north Norfolk, expressed as a percentage of sample number (n), and ratio values. Samples are arranged according to stratigraphic position with the oldest at the base.

Site No	**n**	Triassic white/ colourless quartzite	white/ colourless v. quartz	coloured quartzite	coloured v. quartz	Total qzt+ v. qzt schorl	**Qtz: V. Qzt ratio**	**Coloured: colourless qtz+v. qzt ratio**	Cretaceous flint white	brown	black	Total flint	**Flint: qtz+v. qzt schorl ratio**
(8-16 mm size range)													
UNIT E													
T1/1F	1776	4.7	11.4	5.1	1.9	23.8	**0.74**	**0.43**	30.9	11.9	20.0	62.7	**2.63**
T3/2D	1448	11.1	16.7	4.4	2.2	36.3	**0.82**	**0.24**	20.5	16.9	13.5	50.9	**1.40**
UNIT D													
T2/1C	1081	8.2	19.2	1.5	2.8	32.3	**0.44**	**0.16**	13.4	10.2	31.7	55.3	**1.71**
UNIT B													
T1/1H	278	4.3	12.6	1.4	1.8	21.2	**0.46**	**0.19**	26.3	14.8	24.8	65.8	**3.10**
T1/2G	454	9.9	16.3	3.5	4.4	34.8	**0.64**	**0.30**	17.6	9.3	25.6	52.4	**1.51**
T4/1	423	11.3	14.1	1.7	0.7	27.6	**0.87**	**0.09**	20.9	20.7	14.8	56.7	**2.05**
T5/1	94	8.9	17.9	1.3	2.1	30.9	**0.51**	**0.13**	6.8	31.3	9.8	47.9	**1.55**
(16-32 mm size range)													
UNIT E													
T1/1F	104	9.6	9.6	1.0	0.0	21.2	**1.10**	**0.05**	17.3	15.4	18.3	51.0	**2.40**
T3/2D	289	12.1	8.3	3.8	1.8	26.6	**1.59**	**0.27**	13.2	11.4	14.2	38.8	**1.46**
UNIT D													
T2/1C	63	17.5	9.5	4.8	0.0	33.3	**2.33**	**0.18**	11.1	7.9	23.8	42.9	**1.29**
UNIT B													
T1/1H*	128	10.2	9.4	3.9	2.3	27.3	**1.20**	**0.32**	8.6	8.6	25.0	42.2	**1.55**
T4/1**	144	6.3	10.4	4.2	5.6	27.8	**0.65**	**0.59**	3.5	28.5	9.0	41.0	**1.47**

* includes T1/2G
** includes T5/1

Table 3. Relative pollen abundances in organic units of the 'pre-glacial' sediments at Trimingham, north Norfolk, England.

	Unit B			**Unit D**				
	T2/2C	*T2/2D*	*T3/2C*	*T1/1G*	*T1/2F*	*T3/1F*	*T3/1G*	*T3/2A*
Trees:								
Abies							0.6	
Picea	7.1	5.5	5.3	0.5	5.8	5.7	4.5	5.6
Pinus	29.4	31.5	27.6	26.2	36.1	32.1	23.4	25.6
Ulmus								0.6
Betula	8.2	9.1	19.1	8.3	7.1	13.2	18.8	21.3
Alnus	19.4	7.3	9.2	9.4	9.7	11.3	17.5	10
Carpinus						0.6		0.6
Quercus	1.2	0.6		3.7	1.3	1.3	0.6	1.3
Fraxinus		0.6	0.6			0.6	0.6	0.6
Betulaceae/Corylaceae undiff..	5.3	2.4		0.5	0.6	0.6	1.3	0.6
Coniferales undiff..				1	0.6	0.6	0.6	
Shrubs:								
Ilex								0.6
Ulex type							0.6	
Rosaceae undiff..			0.6	0.5			0.6	0.6
Cylus type	2.9	1.2	0.6	1.6		2.5		
Salix				0.5				
Ericaceae undiff..	4.7	15.8	9.2	3.1	3.2	10.7	11.7	11.9
Herbs:								
Thalictrum			0.6					
Sinapis type				0.5	0.6		0.6	0.6
Caryophyllaceae undiff..			0.6	1.6	1.9	1.9		
Chenopodiaceae	1.2	1.8	0.6	2.6	1.9	1.9		0.6
Filipendula				0.5	0.6	0.6		
Potentilla type	0.6							
Apiaceae undiff..							0.6	0.6
Heracleum type							0.6	
Rumex undiff..					1.3			0.6
c.f. Utricularia			0.6					
Anthemis type						0.6		
Artemisia				1.6	0.6		0.6	0.6
Serratula type							0.6	
Cyperaceae	2.4	7.3	3.3	9.4	5.2	5.7	0.6	3.1
Poaceae	17.6	17.6	22.4	28.3	23.2	10.7	15.6	14.4

Table 3. (continued)

	Unit B T2/2C	T2/2D	T3/2C	Unit D T1/1G	T1/2F	T3/1F	T3/1G	T3/2A
Aquatics:								
Typha latifolia			2	0.5		0.6		
Spores:								
Osmunda regalis							0.5	0.5
Filicales undiff..							0.5	
Polypodium							0.5	
Pteridium	2.1	6		2.9	1.8	2.6	1	1
Dryopteris type	5.2	6	8.9	5.8	4.6	5.6	9.7	9.2
Sphagnum	2.1	6.5	13.9	4.2	6.8	7.7	8.7	12.1
Bryophyte spores				0.8				
Pediastrum	1	2.8	0.5	6.3	13.2	2		
Botrycoccus		1.4		0.4	2.3			
?pre-Quaternary		0.5			0.5			
Tiletia sphagni			1.5			0.5	0.5	
Trilete spores undiff..	1					0.5		
Dinoflagellate cysts		0.6 [1 cyst]		1 [2 cysts]				
Unknown	7	9	4	25	21	17	9	3
Obsured	3	4	4	2		6	1	4
Total Identifiable Land Pollen (excl. spores & aquatics)	170	165	152	191	155	159	154	160

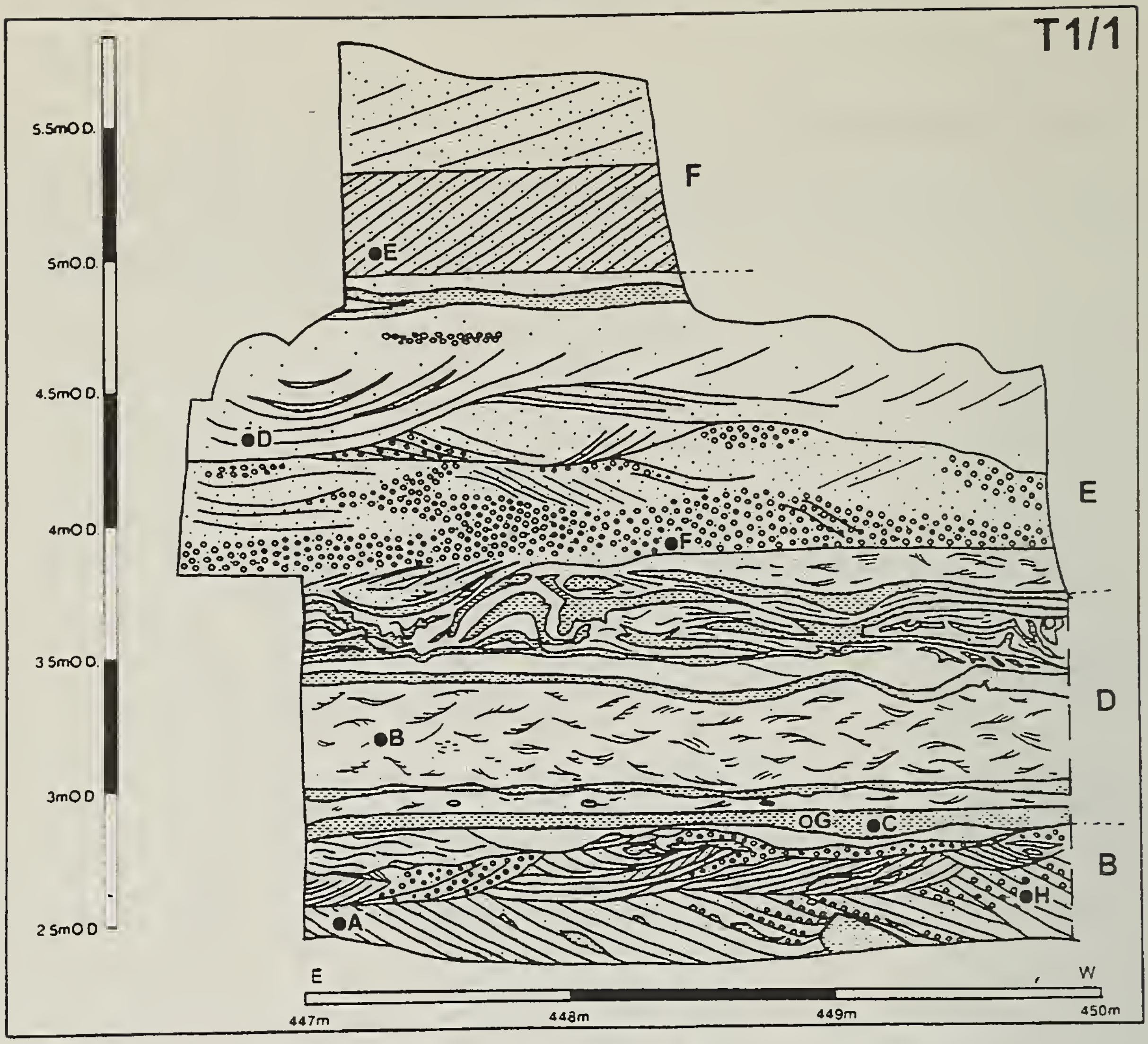

Fig. 2. Sediments and sedimentary structures at section T1/1, Trimingham, north Norfolk, England. The position of this section is located on Fig. 1(c) and the key is given on Fig. 4.

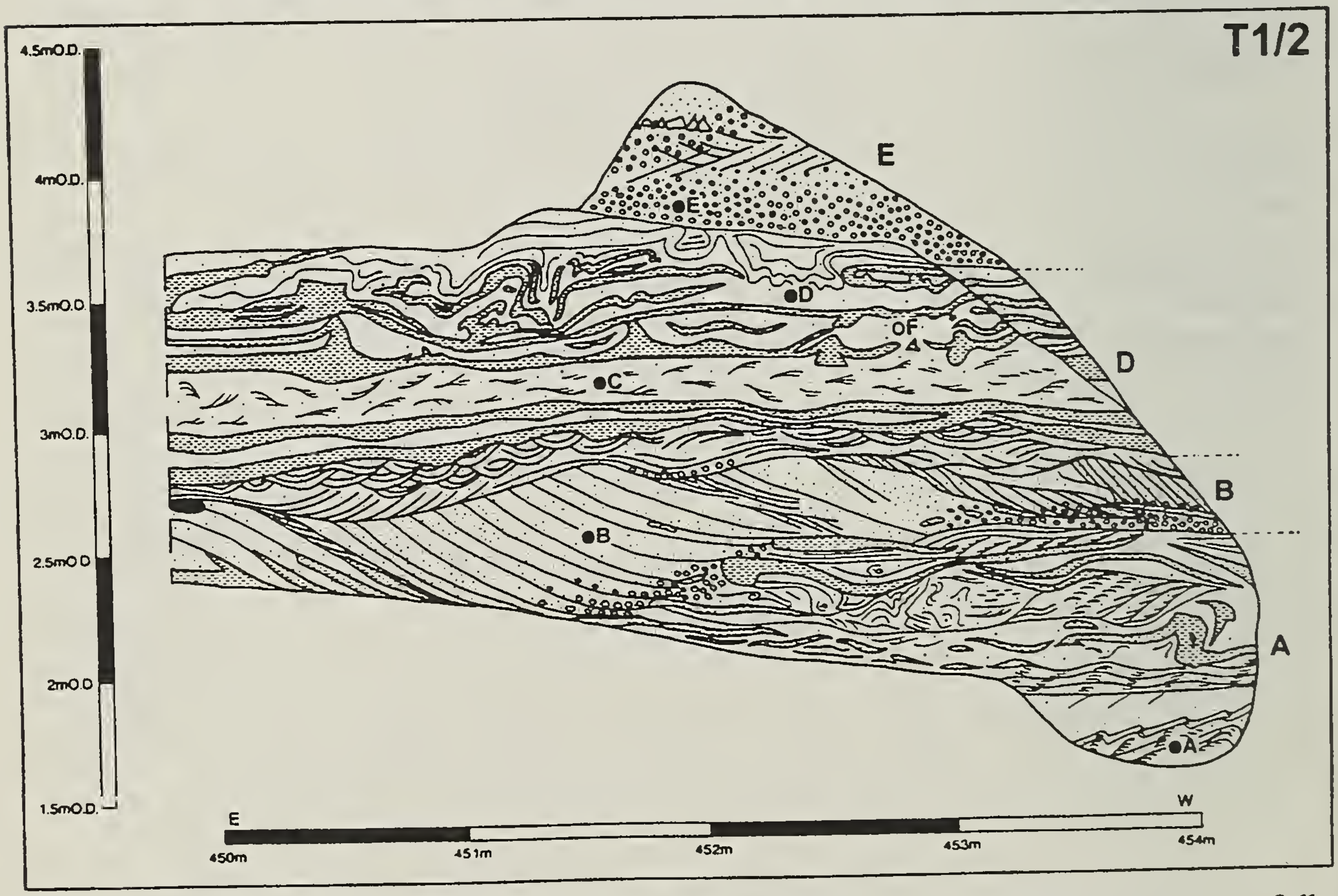

Fig. 3. Sediments and sedimentary structures at section T1/2, Trimingham, north Norfolk, England. The position of this section is located on Fig. 1(c) and the key is given on Fig. 4.

Unit D consists of wavy interbedded fine sands, silts and clays, with a modal particle size of 125-250 μm in the lower part and 63-125 μm in the upper. The erosional contact with Unit B rises slightly from east to west by about 10 cm over 7 m. Two silt and clay beds are overlain by approximately 40 cm of fine sand with discontinuous small ripple bedding, picked out by thin silt cappings. The fine sand is mostly light olive brown (2.5Y5/4), but is olive (5Y5/4) in places. Above this is 50-60 cm of wavy interbedded silts and sands with deformation structures extending throughout the bed. A horizontal erosion surface truncates this unit, having a channel form at the eastern end of the section.

Unit E consists of ~1 m coarse sands and gravels (modal sizes 500-1000 μm and 8-16 mm) that fine upwards to fine sand (modal particle size 125-250 μm) with a wavy silt and clay bed. The coarse sand and gravel is pale yellow (5Y7/3), and the fine sand light yellowish brown to pale olive (2.5Y6/3, 5Y6/4). Poorly defined cross-sets were seen throughout the unit, and the gravels follow the bedding planes. Within the channel-shaped depression to the east of section T1/1, sedimentary structures are picked out by flaser-style silt and clay layers. The transition from coarse to fine within the fining upwards sequence is laterally variable, with a sharp boundary in the east and a more gradual transition in the west. The unit is terminated by an erosional boundary.

Unit F is distinguished by a slight coarsening of grain size (modal size 250-500 μm). It is light grey (2.5Y7/2) with well-defined cross-beds, which are thin in the lower 50 cm, and thicker in the upper 50 cm. The nature of the upper boundary was obscured by slumped till. Palaeocurrent measurements from depositional cross-sets in the lower section cluster between 82° and 158° (ENE-SSE) with a dip ≤26°.

Sections T2/1 and T2/2 (Figs. 4 & 5)

Unit B is visible at the base of section T2/2, and consists of ≤50 cm of pale yellow and yellowish brown (2.5Y7/3 and 10YR5/8) coarse sand cross-sets with a modal particle size of 250-500 μm. The lower contact is obscured by the modern beach, and the upper contact disturbed by cliff collapse. Lenses of laterally discontinuous organic sand are aligned along the cross-sets, some of which include undecomposed pieces of wood. Palaeocurrent measurements on depositional cross-sets cluster between 84° and 110° (ENE-ESE), with dips ≤24°.

Unit D comprises the entire 1 m thickness of section T2/1, although both the upper and lower contacts are obscured. It consists of fine sands (modal size 63-125 μm) interbedded with

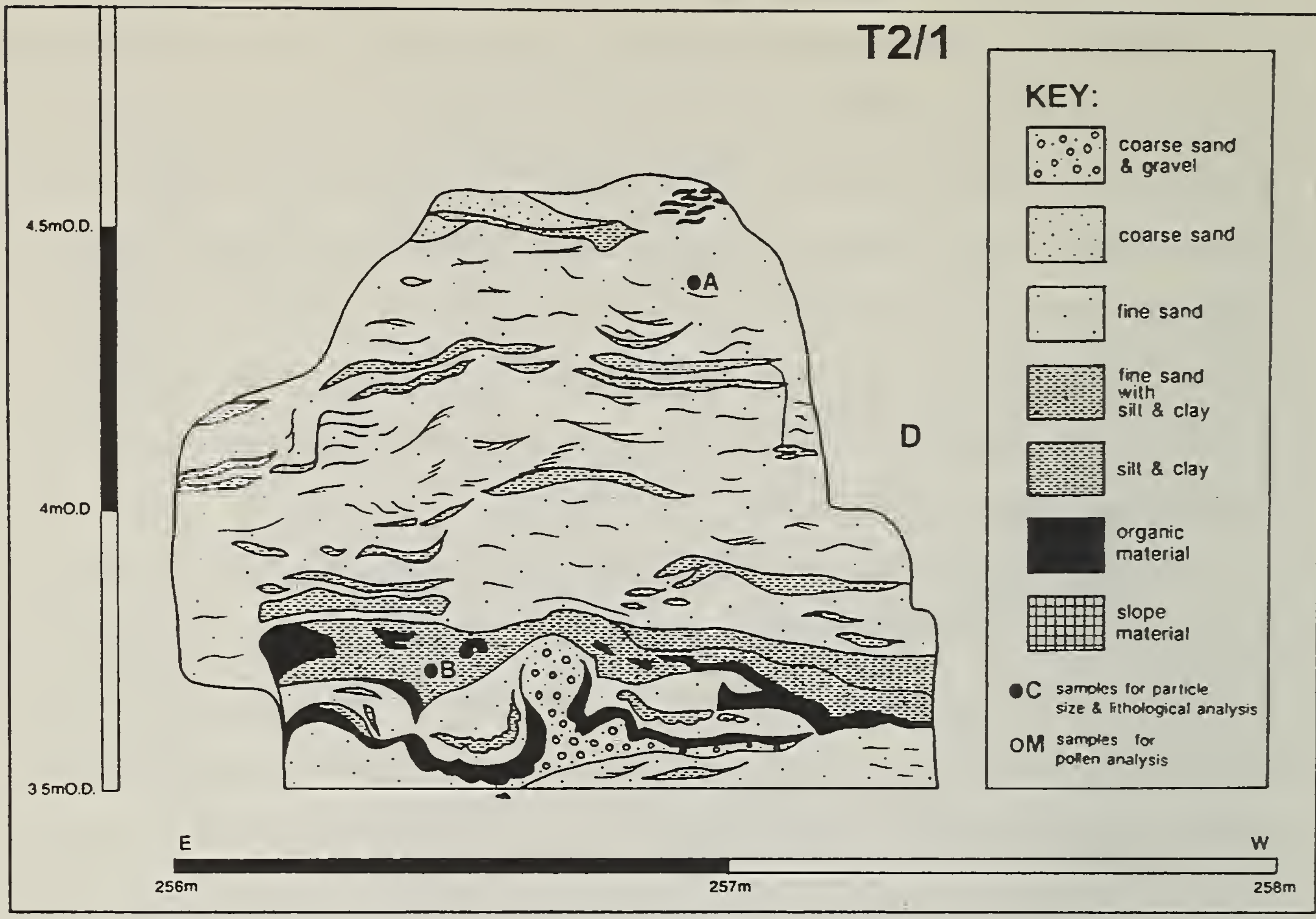

Fig. 4. Sediments and sedimentary structures at section T2/1, Trimingham, north Norfolk, England. The position of this section is located on Fig. 1(c).

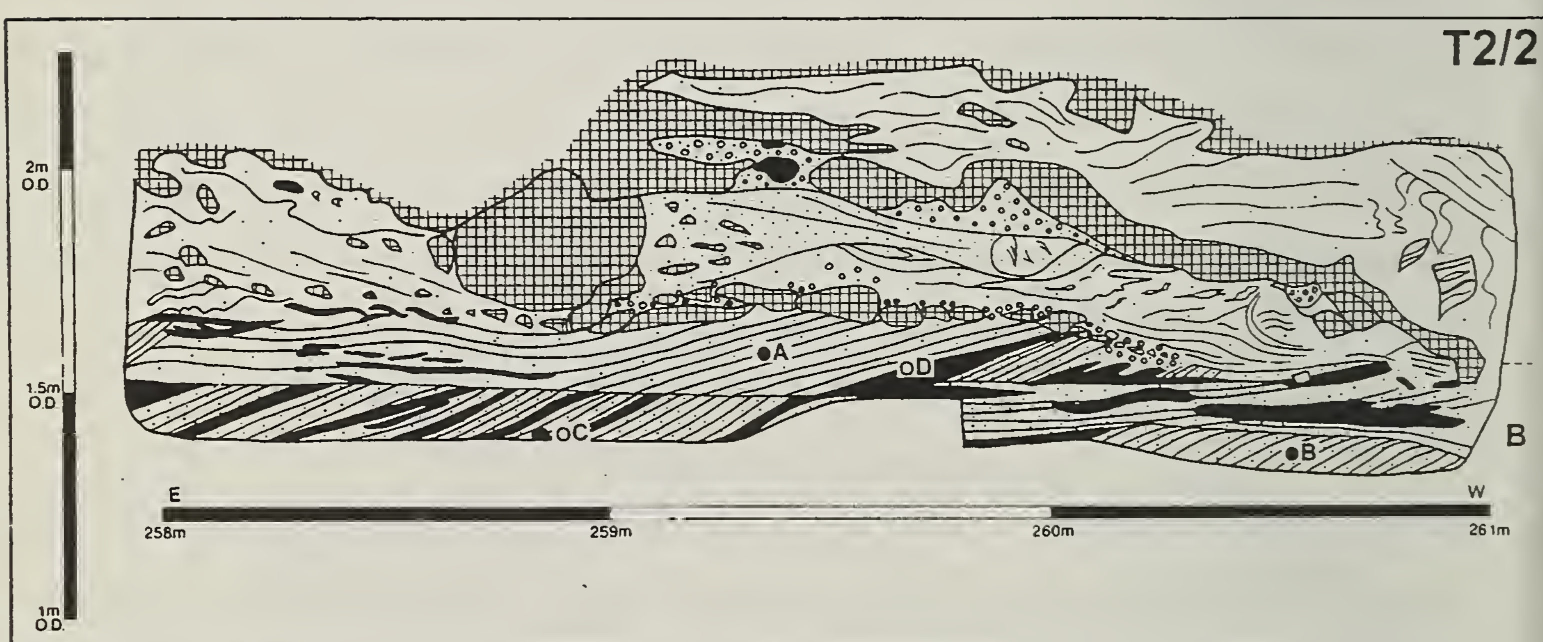

Fig. 5. Sediments and sedimentary structures at section T2/2, Trimingham, north Norfolk, England. The hatched areas represent slumped material. Because of this, and also because of deformation structures and blocks of reworked material in the overlying sands, the area overlyi Unit B is not described or used for interpretation. The position of this section is located on Fig 1(c) and the key is given on Fig. 4.

silts and clays. The sands are pale olive and light olive brown (5Y6/3, 2.5Y5/6). At the base of the section is a distorted fine sand, silt and clay bed with a low organic carbon content (≤2.5%), although locally the unit may be relatively rich in organic material. The lens of coarse sand and fine gravel within this part of the section is pale yellow (2.5Y7/4). Above this, in the main body of the unit, discontinuous small ripple structures are picked out by both thin silt cappings and thicker silt beds (flaser bedding).

Section T3/1 and T3/2 (Figs 6 & 7)

Unit A is exposed for ~3 m in the eastern part of the section (T3/1) and consists of 20 cm of planar bedded fine sands (modal size 125-250 µm) overlain by 20 cm of wavy interbedded fine sands, silts and organic mud. The lower boundary was not seen, and the upper boundary is erosional. The fine sand is light olive brown (2.5Y5/6) and the interbedded material varies from black to grey and dark grey (5Y6/1, N/4).

Unit B is a coarse sand (modal size 500-1000 µm) that varies in thickness between 5 cm and 75 cm, being thicker to the west of T3/2. Colours range between light grey, light yellowish brown, and pale yellow (2.5Y7/2, 2.5Y6/3, 2.5Y8/2, 5Y7/4, 5Y7/2). Structures include discontinuous cross-beds separated by erosional contacts and wavy bedding of clay, silt and fine sand. Sandy organic lenses similar to those in Unit B of T2/2 occur along cross-beds and subhorizontal beds. In places fine gravels also lie along the cross-beds. Palaeocurrent measurements on depositional cross-beds cluster between 95-148° (E-SSE), with dips ≤22°.

Unit C is seen only in the eastern half of section T3/1, and consists of 20 cm thick, sub-horizontal wavy bed of (dark grey, N/4) silt and clay with a modal particle size of 63-125 µm. It rests on an erosional lower contact with Unit B, but after ~3.5 m, it breaks up into wavy beds which appear to interdigitate with Unit B.

Unit D is deformed by two wedge-shaped structures and consequently, is laterally variable in the western half of the section (T3/2). The basal contact is erosional, but the upper contact is obscured. In the eastern half of the section (T3/1) the unit comprises 75 cm of fine sand (modal size 125-250 µm, fining upwards to 63-125 µm) with discontinuous silt-capped small ripples. This is separated into two parts by a thick (<10 cm) discontinuous bed of organic mud. The lower part shows fewer ripple structures and grades upwards from dark greenish grey to light brownish grey (GY4/5, 2.5Y6/2). The upper unit is olive yellow (2.5Y6/6). The lower part of the unit extends only to the middle of the section and the upper part thins to 20-30 cm. In the

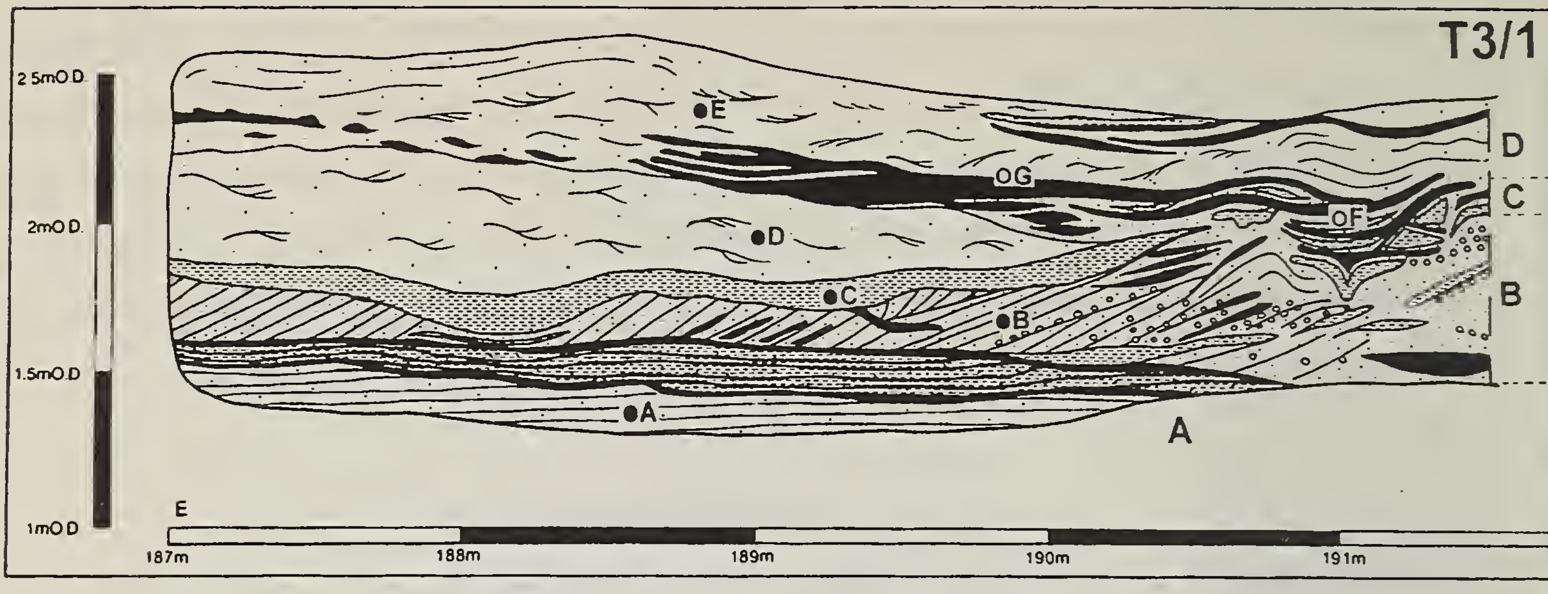

Fig. 6. Sediments and sedimentary structures at section T3/1, Trimingham, north Norfolk, England. The position of this section is located on Fig. 1(c) and the key is given on Fig. 4.

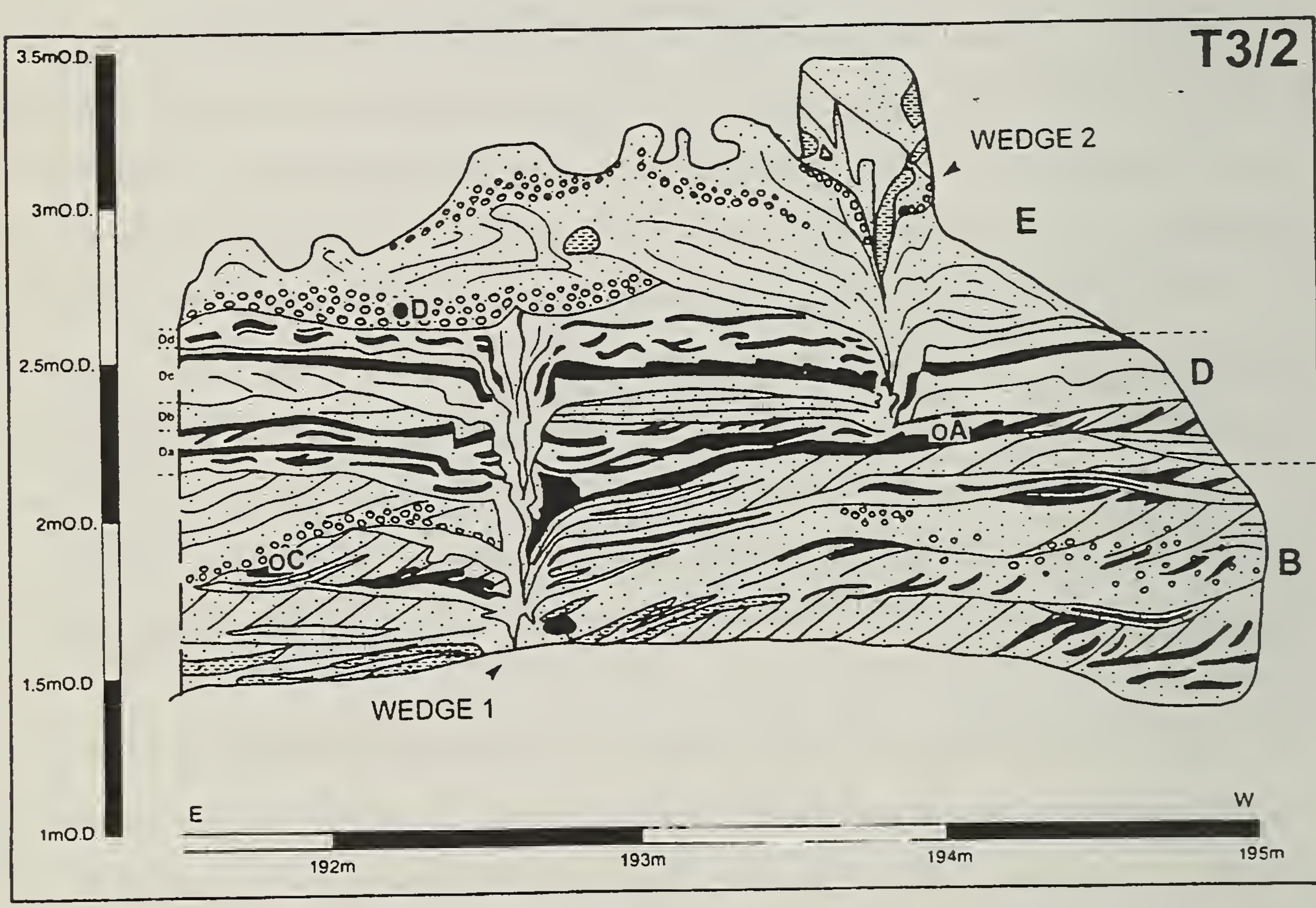

Fig. 7. Sediments and sedimentary structures at section T3/2, Trimingham, north Norfolk, England. The position of this section is located on Fig. 1(c) and the key is given on Fig. 4.

western half of the section (T3/2), the unit is separated into four sub-units (Da-Dd) and consists mainly of fine sand (modal sizes 63-125 µm in Da and 125-250 µm in Dc,), with an intervening bed of coarser sand (sub-unit Db, modal size 250-500 µm). Structures within the unit are disturbed by the lower wedge (wedge 1), but lenses of organic material pick out some sub-horizontal beds and discontinuous small ripples. Sub-unit Da has an organic carbon content of 2.5%.

Unit E has a maximum observed thickness of 50 cm, and overlies unit D with an erosional lower contact. It is seen only in the western half of the section (T3/2). It consists of a sand and gravel layer in the east (modal size 16-32 mm), overlain by a light yellowish and olive brown (2.5Y6/4, 2.5Y5/3) medium-coarse sand containing some gravel (modal size 500-1000 µm). Structures are related to distortion from cliff slumping in the east and the upper wedge (wedge 2) in the west. Fine sand, silt and clay (modal size 63-125 µm) infill the top of wedge 2. The upper boundary is not clear because of cliff slumping, but the wedge appears to be truncated by an erosion surface and buried by medium-coarse sand (modal size 250-500 µm) with some silts and clays.

Section T4 (Fig. 8)

This site was a gap in a zone of extensive cliff-fall in December 1995. It is 4 m long and 2.5 m high.

Unit A consists of 0.25 m of medium, cross-bedded sand (modal size 250-500 µm) with mud drapes on some individual beds overlain by large horizontal beds of sand, silt and clay. All the fine grained sediments have a low organic component, and concentrations of iron and manganese cause local induration at the contact between the sands and the muds. Palaeocurrent measurements from the depositional cross-beds give directions of 110° and 155° (SE) with dips of 25° and 12°. The upper boundary is sharp.

Unit B is the main body of sediment within the section reaching ~1.5 m. It consists of large-scale cross-bedded sands and gravels (modal sizes 250-500 µm and 16-32 mm) in the lower part. These are truncated and overlain by large-scale tangential cross-bedded sands with an amplitude of ~0.40 m and individual lengths of >2.5 m. The upper part of the unit consists of 0.4 m of medium sand (modal size 250-500 µm) with ripple cross-bedding and silt drapes in the ripple troughs (flaser bedding). Palaeocurrent measurements from the large scale depositional

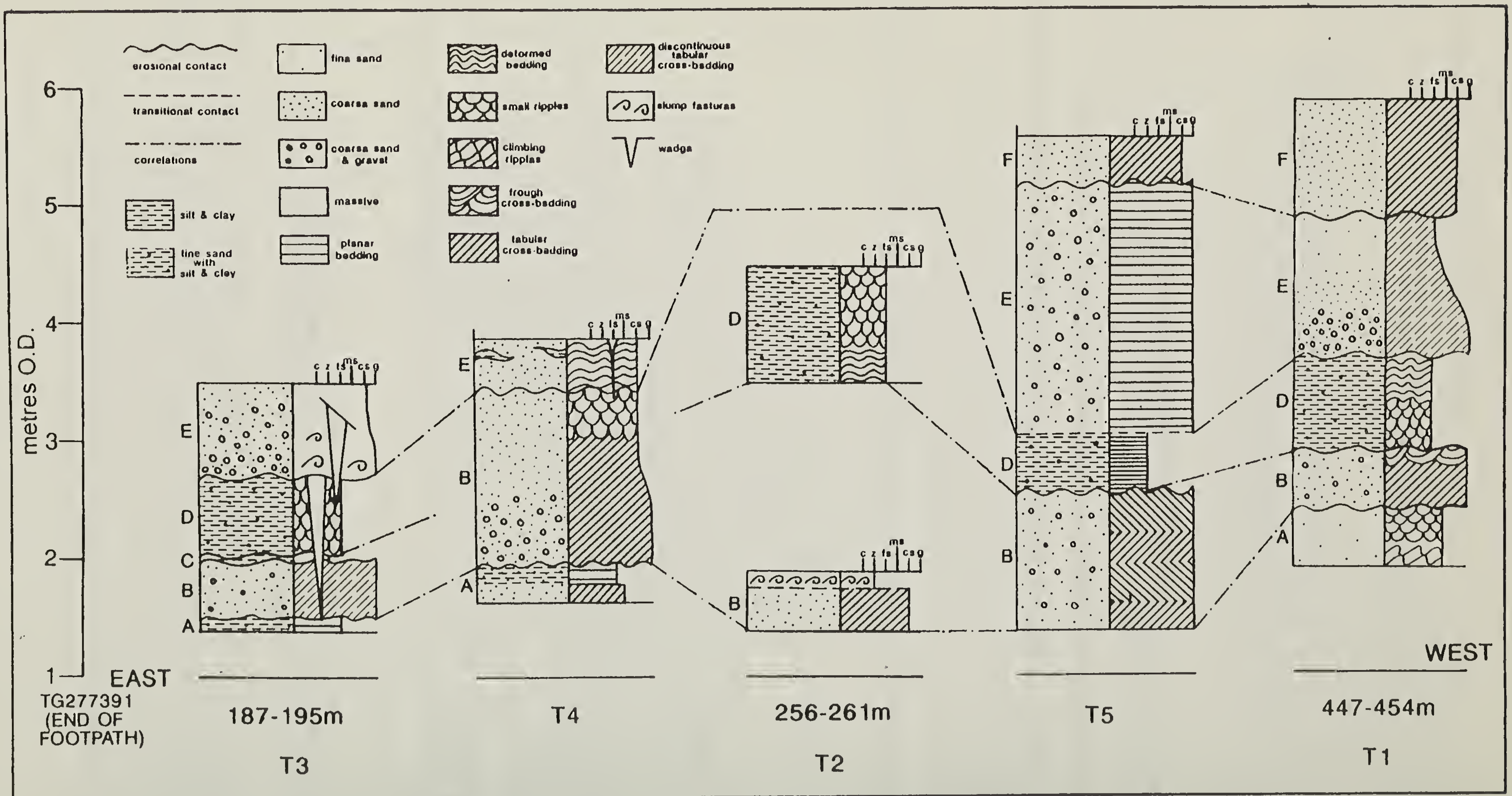

Fig. 8. Correlation of the main sedimentary units in 'pre-glacial' sands at Trimingham, north Norfolk, England. Sediment type is shown at the left side of each log and structures are shown at the right.

beds gives directions between 20° and 135° (NNE-SE) and dips between 7° and 37°. The upper boundary is sharp and truncates the ripple structures.

Unit E is 0.45 m thick and is the highest unit recognised in the section. It consists of a coarse sand (modal size 125-250 μm) with irregular, deformed beds of organic silty clay. This unit is also deformed by a narrow (5 mm) wedge, some 0.45 m in length, that begins at the level of the deformation structures.

Section T5 (Fig. 8)

This site is located at the west end between T1 and T2, and consists of two small exposures, each about 8 m in length with a maximum thickness of ~5 m.

Unit B consists of ~1.2 m of cross-bedded sands and gravels and medium sands showing a herringbone pattern. This is confirmed by palaeocurrent measurement which range from 340-90° (NNW-E) to 150-250° (SSE-SW), with dips in the range of 19-25°. Individual cross-beds reach an amplitude of 35 cms.

Unit D consists of 0.45 m of fine sand with horizontal silt laminations. It rests disconformably upon Unit B.

Unit E consists of 2.2 m of coarse sand with fine gravel forming horizontal beds.

Unit F is exposed in a small part of these sections and shows ~0.4 m of well developed, large-scale cross-bedded coarse sand, resting disconformably upon unit E.

INTERPRETATION OF SEDIMENTARY UNITS

Figure 8 gives a summary of the main units identified at Trimingham with the palaeocurrent directions of Units A and B shown in Figure 9. The correlations indicated in Figure 8 are proposed on the basis of facies similarities. **Unit A** is a fine sand with ripple cross-sets and plane-beds, and interbedded organic silts. Locally there is deformation associated with contrasted textures and the observed palaeoflow direction is to the south-east. **Unit B** is predominantly coarse sand with gravel forming large-scale, often tangential, cross-bedded units. Occasionally there is ripple bedding with clay drapes forming flasers. Organic units are locally interbedded with the sands and these include wood fragments. Palaeoflow directions are towards the north-east and south-east. **Unit C** is of limited extent and is composed of silts and clays. **Unit D** is predominantly fine sand with interbedded silts, clays and organic muds, with silt drapes over ripple laminations. Locally there is deformation associated with contrasting sediment types,

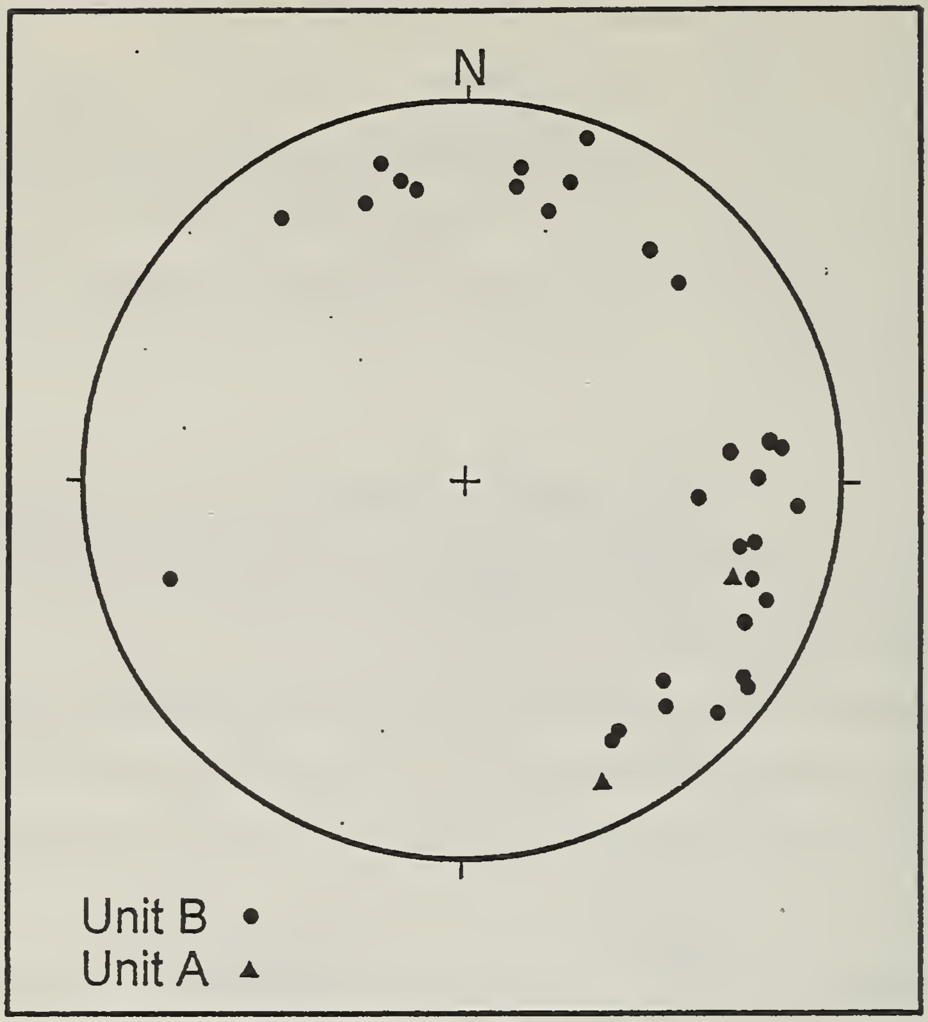

Fig. 9. Dip and orientation of palaeocurrent measurements from 'pre-glacial' sands and gravels exposed at Trimingham, north Norfolk, England. Measurements are taken on the slope of depositional cross-sets of Units A, and B. The measurements on Unit F are not included as it is not clear whether these were formed by the same depositional process.

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but the unit is also disturbed by the wedge structures at T3/2. **Unit E** is a coarse sand with fine gravel. Locally it is interbedded with silts and clays and shows flaser bedding. The two wedges at site T3/2 and the small wedge at T4 formed during the period of deposition of Unit E and extend down into the lower part of Unit D. Where visible, it can be seen that the wedges are overlain by un-disturbed beds of Unit E, but the sedimentary structures in the upper part of the unit are not exposed well enough to determine whether they represent the same depositional environment. **Unit F** is a coarse cross-bedded sand.

Bed-load transportation and saltation processes dominate Units B, E and F (Visher, 1969), and transport by saltation and in suspension is dominant in Units A, C and D. Sedimentary structures indicate flat-bed, ripple and sub-aquatic dune bedforms produced by both low- and

high- regime current flow (Simons *et al.*, 1965), but the silt and clay beds and the silt drapes represent rain-out from suspension in still water. This association of structures, along with the flaser beds, and the bimodal flow directions (Fig. 9) is typical of shallow marine or estuarine tidal environments with the coarser beds transported by current flows and the finer beds by deposition from suspension during slack water (Johnson, 1978). The palaeocurrent trends suggest a coastline aligned roughly north-south controlling the directions of current flow. The changes between sedimentary units probably reflect changes of water depth and position of the tidal-current streams, and it seems likely that higher energy conditions were most important at this site in view of the dominance of coarse grained sediment and the absence of bioturbation. Deformation structures are related to contrasted sediment textures and appear to be load structures caused by inverse density gradients in sediments with very high pore-water pressures. Although there can be rapid facies changes in offshore environments both in time and space, it is considered that in view of the repeated sequence across the study area the units identified are time equivalent.

POLLEN ANALYSIS

Description

Pollen analysis was carried out on 7 samples from organic beds in Units B and D from sections T1/1, T1/2, T2/2, T3/1 and T3/2, and the relative abundances of pollen types are listed in Table 3. The location of the sample points is shown on the appropriate section drawings. The most abundant tree types are *Pinus*, *Betula* and *Alnu*s, with Ericaceae shrubs. Herbs are dominated by Poaceae, but also include Cyperaceae, Caryophyllaceae and Chenopodiaceae. Also present are significant numbers of spores, such as *Pteridium*, *Dryopteris* type and *Sphagnum*. Terrestrial woody detritus is also found, but has not been identified to species level.

Interpretation

The taphonomy of the samples is likely to be complex since the host materials are associated with current flow and thus pollen may be representative of an entire catchment, rather than local vegetation only (Muller, 1959; Groot, 1966; Heusser, 1988). However, Heusser (1988) found that local coastal vegetation is over-represented in the nearshore shelf sediments, whereas regionally representative assemblages characterise mid-slope sediments. West (1980) identified local stream flow patterns as another source of bias, and emphasized that smaller and more

buoyant pollen grains are likely to be selectively transported greater distances. Likewise, pollen preservation is a significant issue. Broken, crumpled and degraded grains create problems of identification and hence the possibly of under-representation of fragile taxa. Within the Trimingham samples, highly resistant grains such as *Pinus*, *Alnus* and *Betula* are common, whereas less resistant grains of taxa such as *Carpinus*, *Ulmus*, *Quercus*, *Fraxinus* and *Salix* are less common. This is not conclusive evidence for bias as resistant pollen types such as *Polypodium*, *Tilia*, and *Corylus* are also absent or poorly represented, but it does re-enforce the need for caution when interpreting the pollen data at Trimingham.

The pollen analysis provides further evidence for a marine depositional environment in the occurrence of dinoflagellate cysts (Edwards, 1993) and the presence of Chenopodiaceae which may indicate salt-marsh conditions. The dinoflagellate cysts included within the count and are assigned to *Operculodinium centrocarpum* on the basis of comparison with the description in Reid (1974). These indicate non-specific marine conditions that could be found from the Caribbean and the coast of Africa throughout the North Atlantic to north of the Arctic Circle (Harland, 1983).

However, the pollen is mainly from fresh-water and terrestrial taxa. *Typha latifolia* is the only aquatic noted, and that rarely, but both *Pediastrum* and *Botrycoccus* were found in greater abundance, indicating the presence of fresh-water. Cyperaceae, *Sphagnum*, *Pteridium*, *Rumex*, *Salix* and *Alnus* also favour damp ground. However, several dryland species, including *Betula, Pinus, Picea, Corylus*, Ericaceae, and *Ilex* were also included in the count.

This combination of species suggests, with the caveat of taphonomy outlined above, the proximity of a temperate alder-carr landscape. This landscape would be a mosaic ranging from wetter areas with sedges and damp-loving plants, possibly even with some peat formation (*Sphagnum*), and fringed with alder (*Alnus*); to a drier open landscape with *Betula* and *Pinus* woodland, Ericaceae and *Corylus* shrubs, grasses and herbs. This is likely to represent a local vegetation community typical of Heusser's (1988) nearshore domain.

CLAST LITHOLOGY

Description

Table 2 shows the percentage lithological composition of the gravel fraction of the sands and gravels. The major component of the gravels is flint (8-16 mm = 47.9-68.7%, 16-32 mm = 41.0-51.0%) followed by quartzite, vein quartz and schorl (8-16 mm = 21.3-36.3%, 16-32 mm = 21.2-

33.3%). Chattermarked flints also form a significant group (8-16 mm = 4.7-11.4%, 16-32 mm = 20.6-28.1%). Indicator lithologies include *Rhaxella* chert (present in all seven 8-16 mm samples, 0.2-3.1%, and in two of the five 16-32 mm samples ≤1.4%), Carboniferous chert (present in all seven 8-16 mm samples, 0.4-7.3%, and in four of the five 16-32 mm samples, ≤2.3%), and Lower Greensand chert (being present in five of the seven 8-16 mm samples, 0.2-1.1%, and in two of the five 16-32 mm samples ≤2.0%). Igneous clasts recorded include acid volcanic rocks and a basalt.

The composition of the sedimentological units is essentially similar, although variations in the frequencies of individual lithologies may suggest that some sources contributed variably. The assemblage does however contrast with the flint dominant gravels described by Hey (1982) that are associated with the earlier Pleistocene gravels (pre pre-Pastonian a) of northern East Anglia. Many of the lithologies can be provenanced with a high degree of confidence (Hey, 1976, 1982; Green and McGregor, 1990; Rose, Gulamali *et al.*, 1996). Quartzite and vein quartz are from the Triassic rocks of west Midland England, Carboniferous chert is from the southern Pennines, *Rhaxella* chert from the southern part of the North York Moors and Greensand chert from the Weald of southern England. Acid volcanic rocks of the kind found here are from north Wales and chattermarked flints are from the Early Pleistocene gravels of East Anglia. Transport of these lithologies from their respective sources is understood in some detail (Hey, 1976, 1980; Hey and Brenchley, 1977; Whiteman and Rose, 1992; Rose, 1994; Rose, Gulamali *et al.*, 1996) (see also Fig. 13 below). The ancestral river Thames, which drained much of southern and Midland England and Wales, is considered the source of bleached quartzite and vein quartz and Greensand chert, along with acid volcanic rocks. The Bytham river, that drained Midland England, is responsible for the transport of coloured quartzite and vein quartz and Carboniferous chert, as well as glauconitic Spilsby sandstone which has not been recorded at Trimingham. A river system called the 'Northern rivers' is considered also to have transported quartzite and vein quartz, Carboniferous chert and *Rhaxella* from the area of the Pennines and North York Moors (Rose, Gulamali *et al.*, 1996, fig. 9).

By consideration of the frequencies shown in Table 2 and expressed graphically in Figure 10 it is clear that the sediments are most similar to the Kesgrave Sands and Gravels and hence are likely to be derived originally from the River Thames drainage basin. However the relatively high frequency of Carboniferous and *Rhaxella* chert suggests a contribution from another source. This is likely to be the 'Northern rivers' as the frequencies of these lithologies are lower than

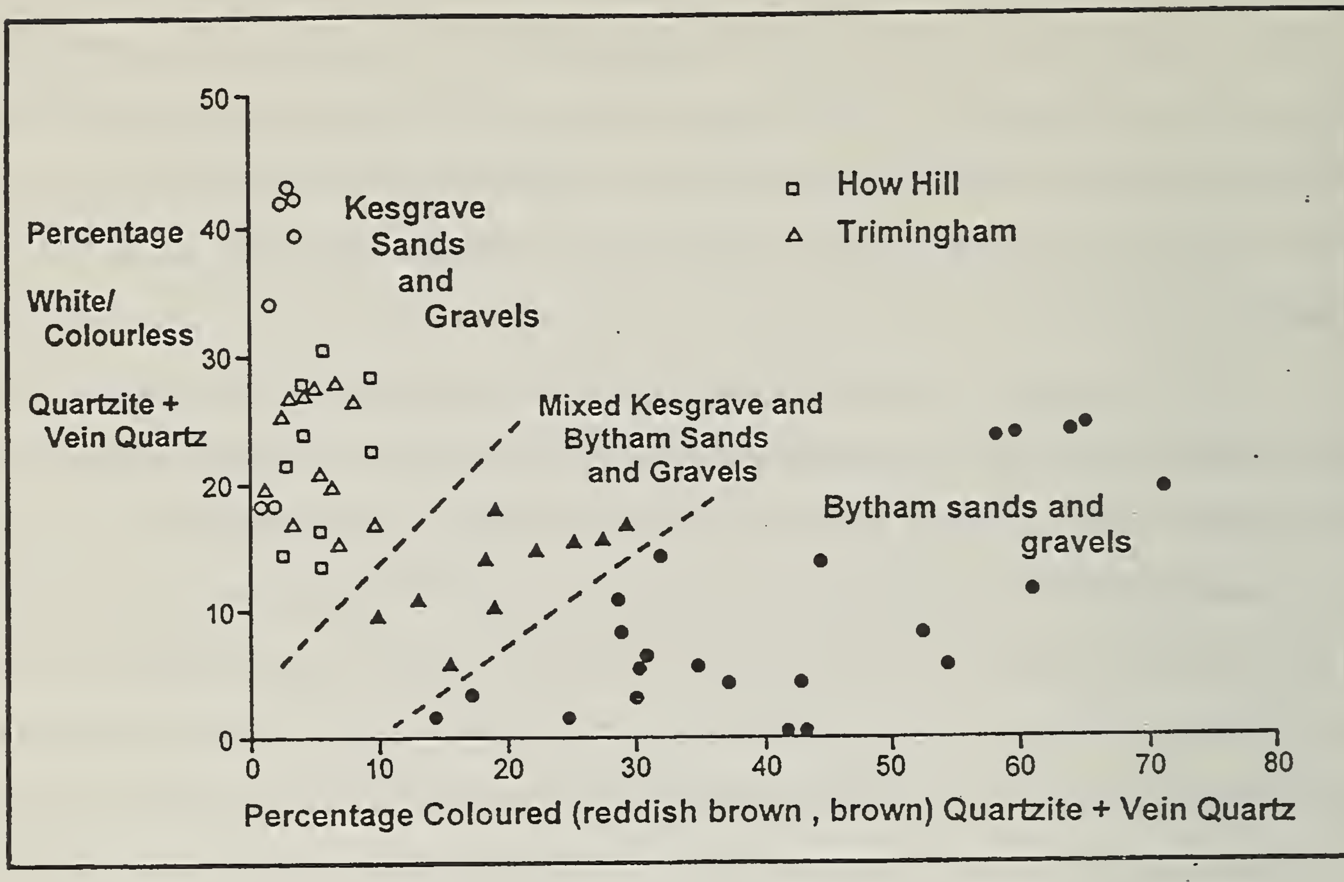

Fig. 10. Relationship between coloured quartzite and vein quartz and colourless quartzite and vein quartz in the deposits at Trimingham compared with samples from the Bytham sands and gravels, Kesgrave Sands and Gravels and mixed Bytham and Kesgrave sands and gravels (Kirby Cane sands and gravels of Leet Hill, see Rose, Lee *et al.* in press). It is clear that the Trimingham deposits are most similar to the Kesgrave deposits. The sites that constitute the Bytham sands and gravels are: Redgrave, Little Fakenham, Feltwell, Shouldham Thorpe, Castle Bytham, Waverley Wood, Knowle and Snitterfield. The sites that constitute the Kesgrave Sands and Gravels are Cutmaple, Gosfield, Badwell Ash, and How Hill. Samples from both 8-16 mm and 16-32 mm size ranges are included. The samples from How Hill, that are closest to Trimingham, are indicated separately.

those typically found within the Bytham sands and gravels (Rose, Lee *et al.*, in press) and the deposits at Trimingham do not have the high frequencies of coloured quartzite and vein quartz of the Bytham sands and gravels (Fig. 10). The samples are most similar to 'Type C' gravels of Green and McGregor (1990) who also suggest a provenance from the Midlands and Pennines.

WEDGE STRUCTURES

Significant numbers of wedges have previously been recorded along the north Norfolk coast (West, 1980; Rose *et al.*, 1985) including individual features recorded within the area of study (West, 1980, Fig. 29; Fish *et al.*, 1998). Although Worsley (1996) has cautioned against attribution of wedge structures to periglacial processes without careful study, and Rose, Gulamali *et al.* (1996) have shown that wedge-shaped features can be formed by water escape from sediments, many have characteristics diagnostic of permafrost conditions (Fish *et al.*, 1996) and have been shown to be part of the Barham Soil (Rose *et al.*, 1985) which formed on ephemeral aggrading land surfaces during the late Cromerian and early Anglian prior to burial by Anglian ice.

Wedge 1 (Fig. 11a)

Wedge 1 of T3/2 is developed in interbedded fine-medium and medium-coarse sands of Unit D, the upper part being truncated by the erosion at the top of this unit. It is >1m long, and <25 cm wide at the top. It consists of a vertical body of moderately-sorted fine and very fine sand with discontinuous vertical laminations. Normal faulting parallels the wedge at the top. In the lower part of the wedge, there is some up-turning of strata in addition to down-turning and slump structures. The wedge was not excavated because of danger of cliff-fall, but during the study period cliff erosion showed that it extended back at least 50 cm into the cliff as part of a linear feature. Particle size analysis (Table 1; Fig. 12) shows that the infill sediment is different from the host materials indicating that it has been derived from beyond the site. On the basis of the derived infill, up-turned structures at the margin and linear form, this feature is interpreted as a permafrost thermal-contraction-fissure, probably formed as part of polygonal, non-sorted patterned-ground that was infilled by wind-blown sediment, before collapsing during the melting of permafrost (Black, 1976; Harry and Gozdzik, 1988). Thus it is like the feature studied by Fish *et al.*, (1998) and is considered a composite wedge. The relationship of the wedge to Unit D suggests that it is epigenetic to this sediment body (French, 1996).

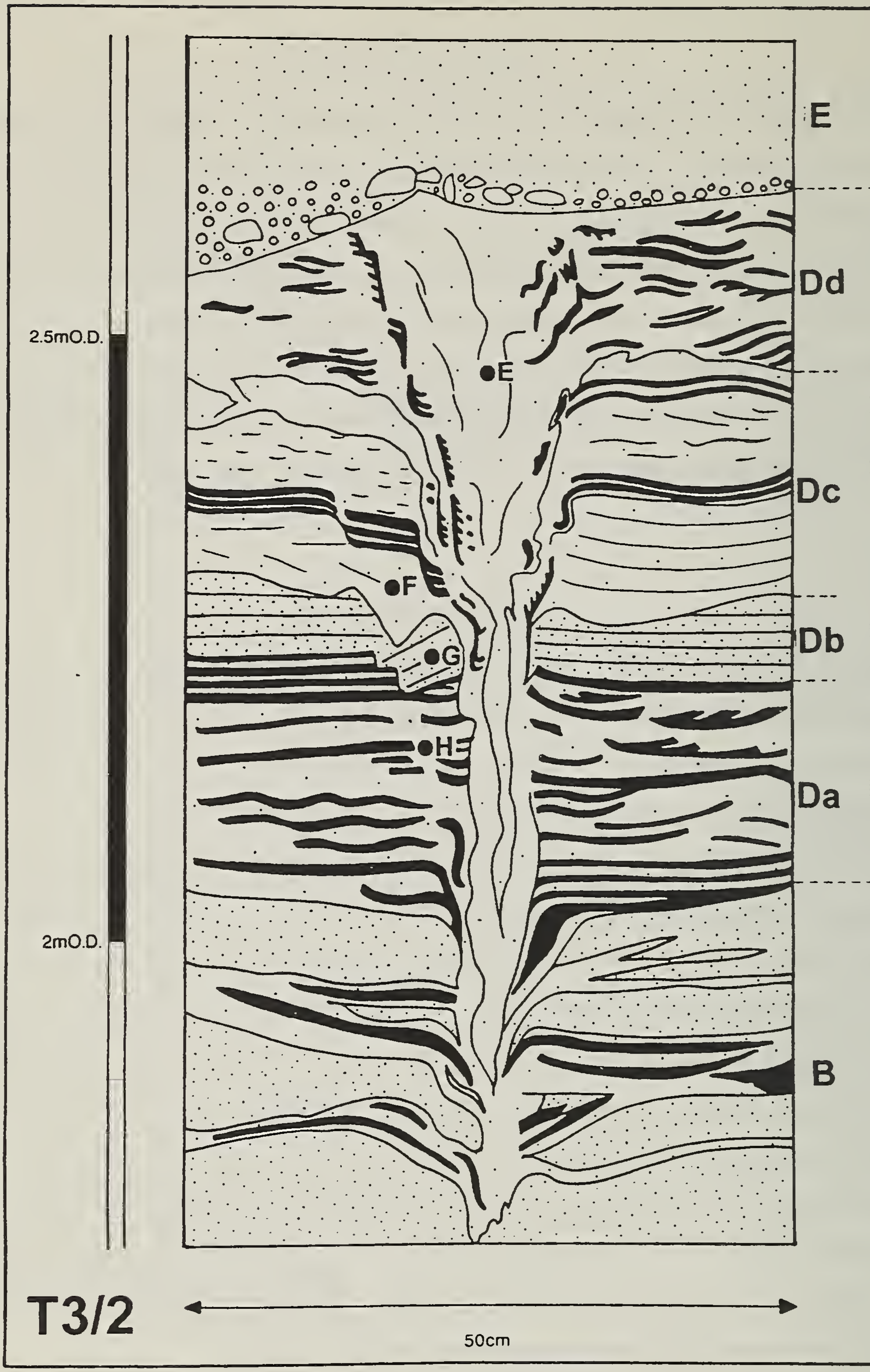

Fig. 11a.. Wedge 1 from section T3/2 of the pre-glacial sands at Trimingham, North Norfolk, England. The key to symbols is shown in Fig 4.

Wedge 2 (Fig. 11b)

Wedge 2 is developed mostly in coarse sands of Unit E extending down into fine sands of Unit D. It is 1m long and <50 cm wide. The wedge infill is hard to distinguish from the host sediments, and consists of vertical lineations of fine grained sediment within the coarse sand host material. Strata at the sides of the wedge are downturned and distorted, but can be traced either side of the structure, although higher on the right than on the left. No upturning of strata is seen. The top of this wedge is truncated diagonally and obscured by cliff fall.

On the basis of this evidence there are no diagnostic features that indicate a permafrost origin, and following the caution recommended by Worsley (1996) this wedge is not interpreted as periglacial. It is suggested that it formed due to desiccation of the fine sediments of Unit D, or indeed by water-escape from these same materials in a similar fashion to the structures recorded at West Runton and How Hill (Worsley, 1996; Rose, Gulamali *et al.*, 1996).

Small wedge - site T4

This is developed in Unit E at site T4 with a length of 0.45 m and a width of <5 mm. There is little evidence to indicate the genesis of this feature, but its position in coarse sand with fine gravel suggests that it is unlikely to be a desiccation or water escape structure and is typical of a frost crack, developed in permafrost regions by limited thermal contraction (French, 1996).

Palaeoenvironmental significance of the wedges

The presence of a periglacial composite-wedge cast extending into Unit D, probably from Unit E, and the frost crack in Unit E shows that a land surface was exposed to surface processes for sufficient length of time for thermal contraction to take place and non-sorted polygonal patterned ground to develop. This means that Unit E must have been dry land while mean annual temperatures were in the order of -1°C to -11°C (Black, 1976; Harry and Gozdzik, 1988).

DISCUSSION

The bulk of the sorted sediments beneath the glaciogenic deposits at Trimingham are interpreted as having formed in a shallow marine environment with changes of water depth, changes in the direction of current flow and variations in the proximity to the contemporary coastline. Unit A suggests low energy sand and silt transport probably in shallow, marginal creek conditions. Unit B however, suggests much greater water depths with powerful tidal currents running parallel

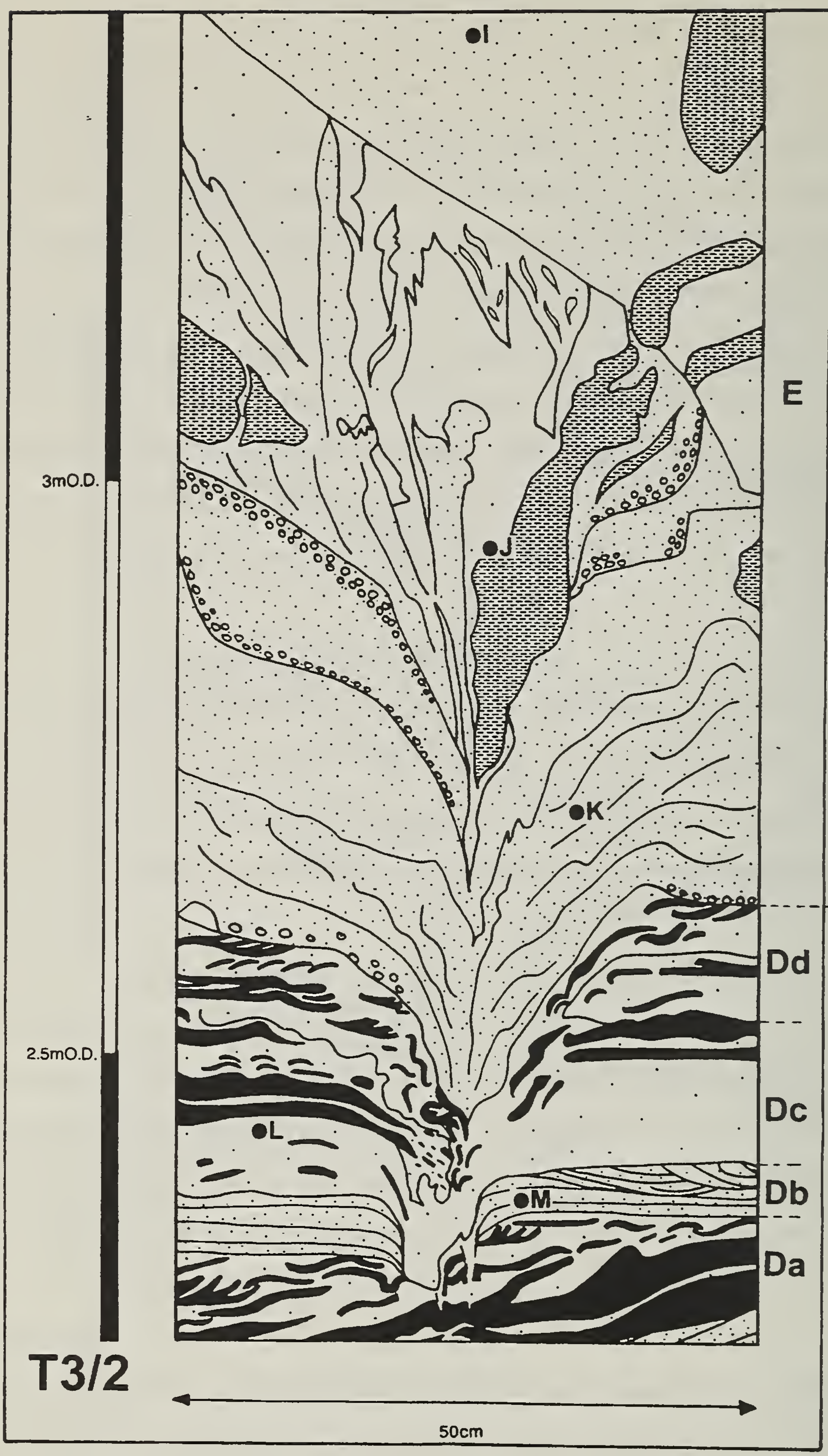

Fig. 11b. Wedge 2 from section T3/2 of the pre-glacial sands at Trimingham, North Norfolk, England. The key to symbols is shown in Fig 4.

with a roughly north-south trending coastline, and forming sub-aquatic dunes and tidal current channels and bars. Occasionally this unit also shows signs of lower flow-regimes, probably due to shifts in the position the tidal current channels. It is likely that rapid deposition of this unit caused the load deformation of the water saturated sediments in the underlying Unit A. Unit C indicates a return to lower energy, probably shallow water conditions. Unit D is the product of classical, moderate energy tidal-flow sedimentation (Reineck and Singh, 1980), with sands, silts and clays and local concentrations of organic detritus within the mineral sediments. There is no evidence of the sediments drying-out, and the load deformation structures suggest that very high pore-water pressures accompanied sedimentation. In contrast, Unit E shows highly variable water depths, with evidence of sand and gravel transport by relatively powerful, deep-water tidal currents, interrupted by periods when sand bars remained at the surface and were disturbed by thermal contraction cracking in a permafrost environment. The upper part of Unit E and Unit F reflect further submergence with rapid deposition, possibly causing loading and the development of the water escape structure of Wedge 2.

The pollen evidence from the lower organic deposits suggests that sedimentation initially took place in a temperate climate and the presence of fragile pollen and spores confirms the presence of closely adjacent land areas. The pollen assemblage suggests that these land areas were vegetated by alder-carr, and by reference to the regional study of West (1980) it is likely that the sediments at Trimingham were deposited in the nearshore domain. In contrast, the Cromer Forest-bed *sensu-stricto,* represented by sediments elsewhere along the coast at localities such as West Runton and Corton, formed on the adjacent floodplain. This freshwater floodplain landscape ranged from open river channels, to backswamp areas with sedges and alder (in which the forest-bed *sensu-stricto* was accumulated), to a drier land with birch and pine woodland amongst heath and hazel, grasses and herbs.

However, the upper ice-wedge casts from this and other studies, and the cold climate pollen recorded by West (1980) suggest that the climate changed from temperate to permafrost. It is not possible to determine whether the uppermost sediments in Unit E and the sediments in Unit F are marine or fluviatile and therefore, it is impossible to say whether relative sea-level remained the same or experienced a fall as would be expected from a climatic deterioration.

These sediments were deposited as part of the deltaic environment that stretched across the southern North Sea (Fig. 13) (Funnell, 1996). This delta was formed by major river systems such as the Rhine, Meuse and 'Baltic' rivers from Europe and the Thames, Bytham and 'Northern

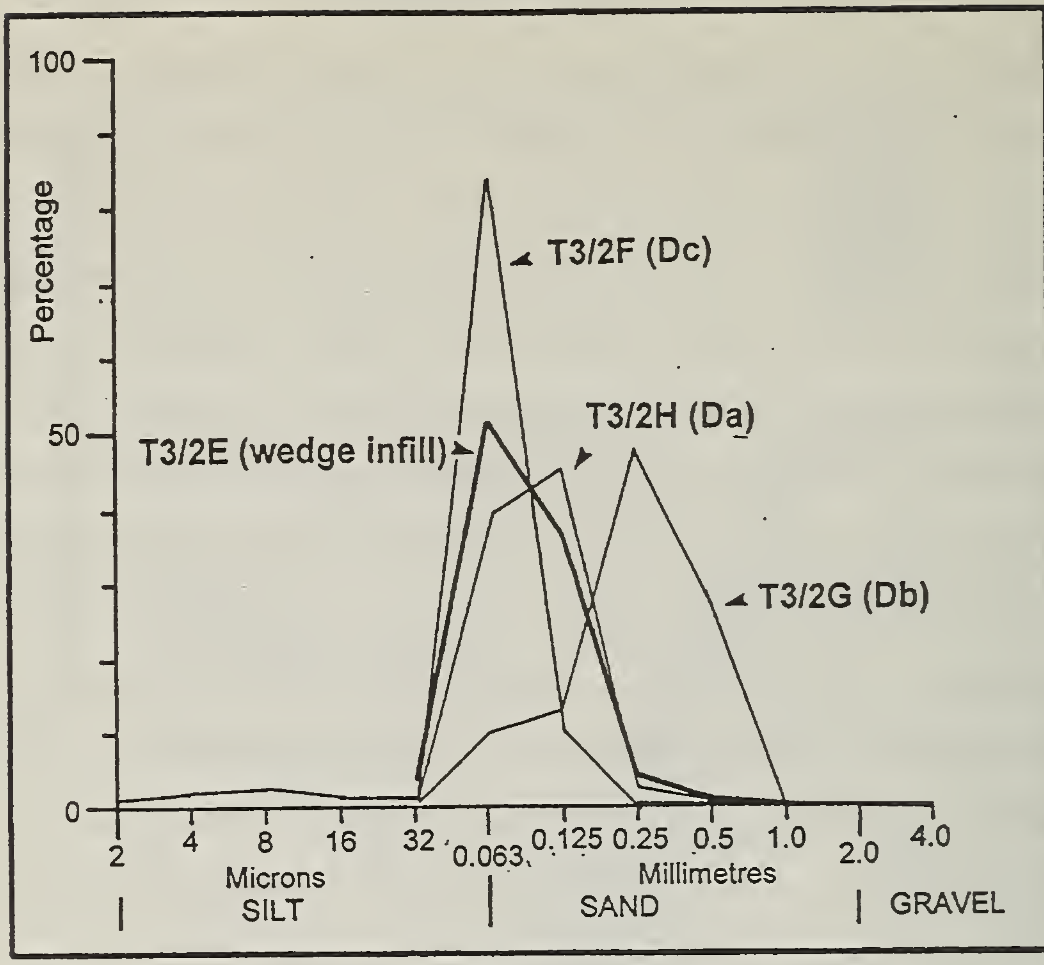

Fig. 12. Particle size properties of sediment infilling Wedge 1 and the host materials. It is clear that there is no relationship between the infill and these host deposits.

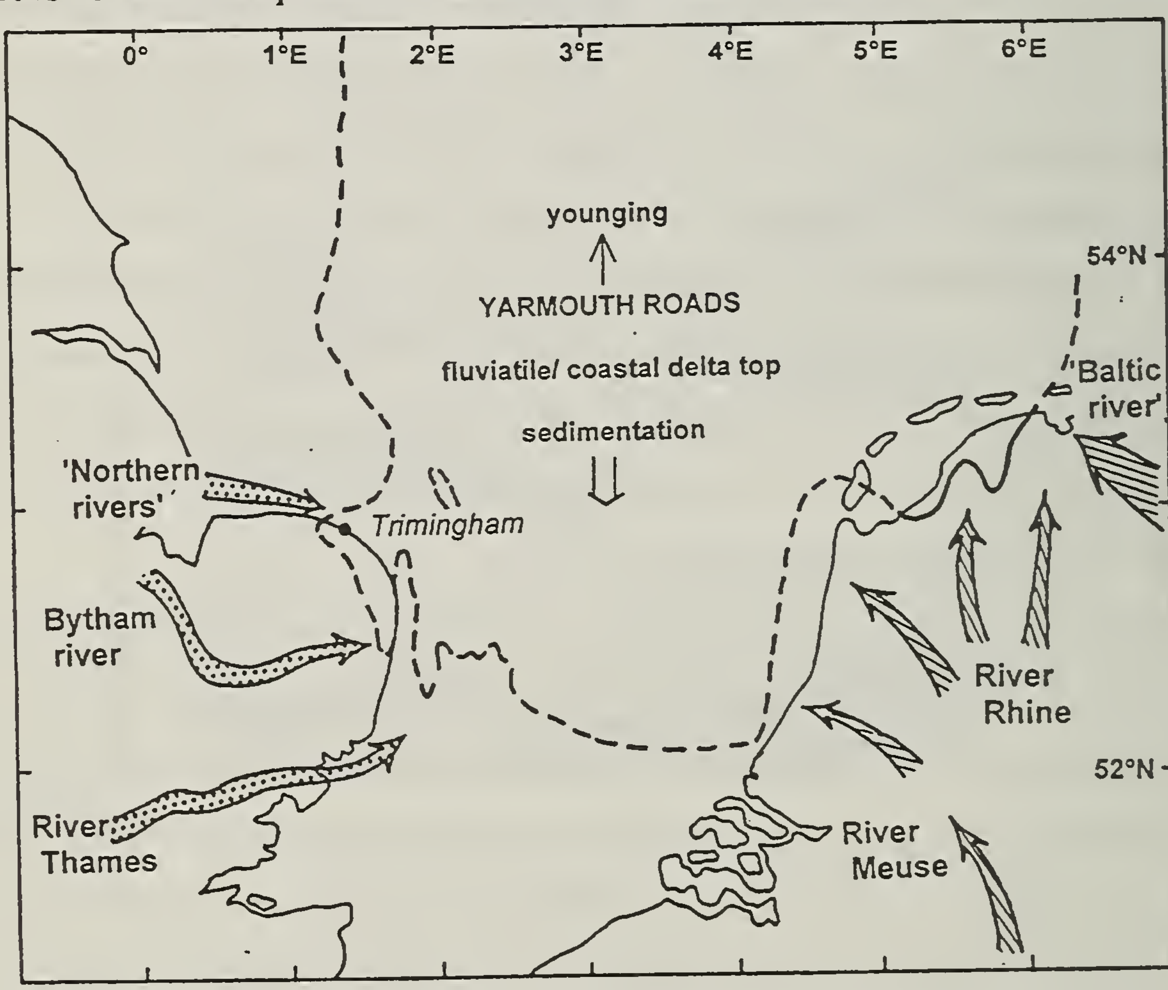

Fig. 13. Palaeogeography of the southern North Sea region at the time of formation of the pre-glacial sands and gravels at Trimingham (Based on Funnell, 1996).

rivers' in eastern England. Further from the coast, deposition is represented by the Yarmouth Roads Formation of the southern North Sea (Cameron *et al.*, 1992). The conditions that existed at this time were such that the river systems were transporting coarse sediments into the coastal system, unlike earlier in the Quaternary when only fine grained material was transported and coastal sediments were composed primarily of local material (Rose, 1994)

CONCLUSIONS

1) The sorted sediments beneath the glaciogenic deposits at Trimingham are part of the Cromer Forest-bed Formation and include elements of the Barham Soil formed on ephemeral land surfaces of an aggrading depositional system. These probably date from the a later part of the Cromerian Stage, given the temperate pollen assemblage in the lower part of the deposits, through into the periglacial climate of the early Anglian Stage.

2) The environment of deposition of the pre-glacial sediments at Trimingham was shallow marine, dominated by tidal current processes with local slack water sedimentation. This interpretation is based on the presence of interbedded sands and silts, small-scale ripples with silt drapes, bimodal palaeocurrent directions, laterally extensive coarse sand cross-sets, a mixed lithological assemblage, and dinoflagellate cysts. Other sites retain a marine molluscan fauna. This dynamic environment is represented by wide variations in water depth, and changes of tidal current direction and position. Fragile freshwater algae, pollen and spores within the sediments, which would survive only limited transport, suggest that the site was close to the shore.

3) Lithologies within the gravels indicate provenance from the ancestral Thames (Kesgrave Sands and Gravels with high quartzite and vein quartz content, Lower Greensand chert and Welsh volcanic rocks) and from as yet uncharted 'Northern rivers' (with Carboniferous and *Rhaxella* cherts). In this respect they are similar to shallow marine deposits found at How Hill, north-east of Norwich (Rose, Gulamali *et al.*, 1996).

4) The pollen found in the organic lenses suggests that these deposits formed adjacent to a temperate-climate floodplain with alder-carr backswamp areas grading into drier, more open birch and pine woodland with shrub and dryland herbs.

5) Environmental conditions at the site concluded with permafrost forming on exposed bar surfaces. Whether the sediments that bury these wedge patterns are marine or fluviatile is not resolved.

6) In terms of the current controversy about the extent of the Thames river deposits within northern East Anglia, this work suggests that the river Thames did not reach the area of the present north Norfolk coast, but contributed sediment to a coastal domain of which this site is a part.

ACKNOWLEDGEMENTS

The work for this study was carried out whilst RMB was a Quaternary Science MSc. student in the Department of Geography at Royal Holloway, University of London and in receipt of NERC Studentship GTO3/97/131A/ES. Warmest thanks are expressed to John Briant for assistance with the GTS and to Professors Brian Funnell and Richard West for observations arising from their reviews of the paper.

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CONTENTS

The Geological Society of Norfolk exists to promote the study and understanding of geology, particularly in East Anglia, and holds monthly meetings throughout the year.

Visitors are welcome to attend meetings and may apply for membership of the society. For further details write to The Secretary, Geological Society of Norfolk, Castle Museum, Norwich NR1 3JU.

Copies of the Bulletin may be obtained from the Secretary at the address given above; it is issued free to members.

The illustration on the front cover is figure 4 from the article by Boomer and Woodcock in this issue of the Bulletin. It shows a model of barrier drowning and subsequent landward migration during enhanced relative sea-level rise that might be applicable to the formation of the Stiffkey Meals. **A)** shows position of the original offshore barrier with the approximate position of the high water mark (HWM) at that time; **B)** shows the position of the new barrier and new HWM.

ISSN0143-9286

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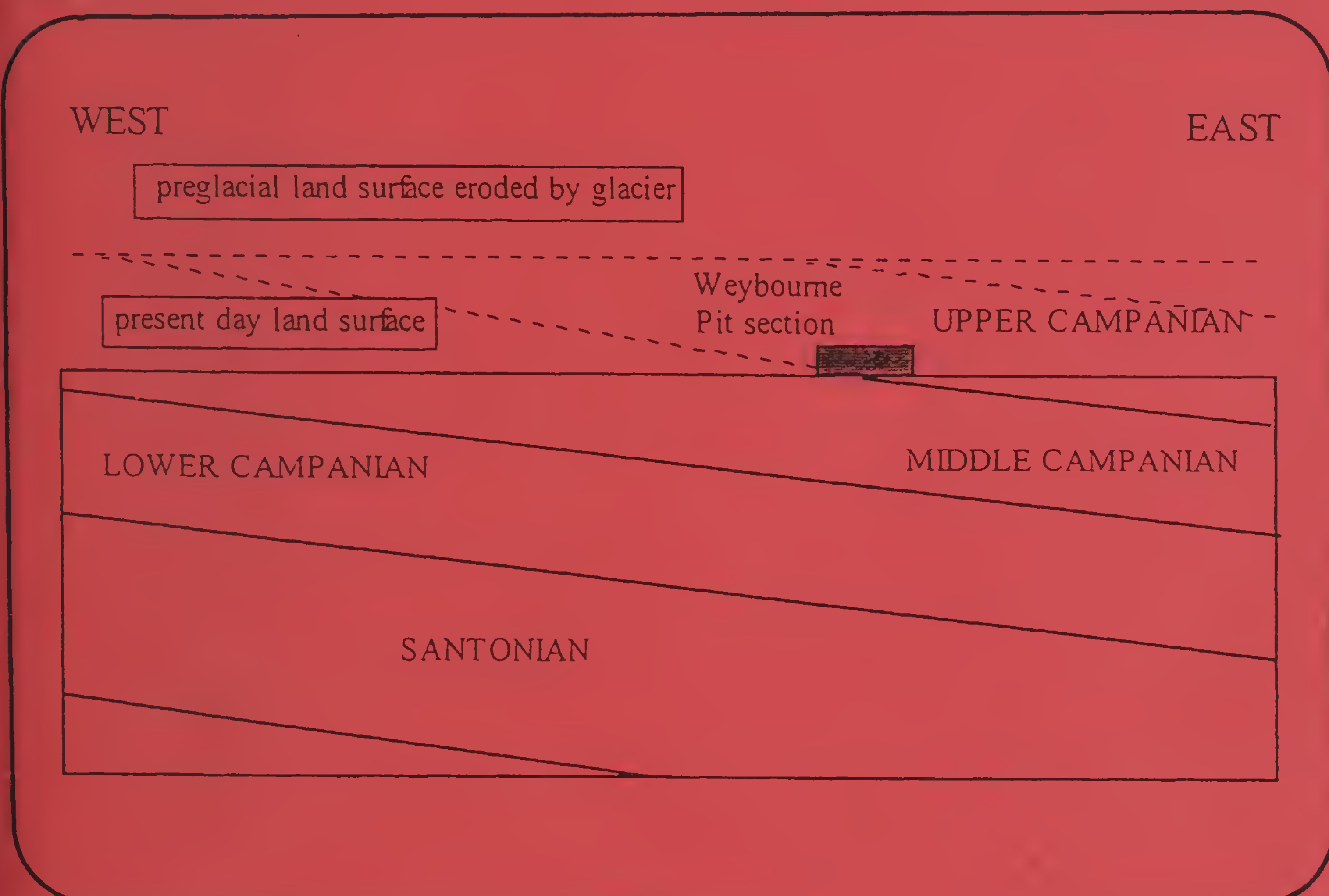
BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

(FOR ARTICLES ON THE GEOLOGY OF EAST ANGLIA)

NO.50 2000



PUBLISHED 2000

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Glacial geology and provenance at Weybourne

East Anglian erosion

BULLETIN OF THE GEOLOGICAL SOCIETY OF NORFOLK

No. 50 (2000) Published 2000

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EDITORIAL

Bulletin No. 50 contains three papers that cover related aspects of Quaternary geology in East Anglia. The papers by Fish et al., and Banham, although submitted separately, both detail till provenance in north Norfolk. The former concentrates on a new way of provenancing tills based on Chalk microfossil content, while the latter is essentially a data archive of information on Quaternary gravels collected by the late Colin Ransom. The paper by Clayton is a welcome review of some of his recent work on quantifying Quaternary erosion by various agents. This paper summarizes the excellent presidential address given by Prof. Clayton to the society in 1998.

Health of the Bulletin

The last two Bulletins (49 and 50) have appeared on time with the publication schedule completely up to date. It is therefore a good time to assess the relative health of the Bulletin in terms of article submissions and number of pages published. I took over full editorship of the Bulletin in 1993, and upon publication of Bulletin 42 (for 1992 but

eventually published in 1995), it was clear that the low rate of article submissions was not going to sustain the 100 page Bulletins that previous editors had strived to produce. In my editorial in Bulletin 42, I announced my intention to publish a succession of 50 page Bulletins to catch up on the publication schedule. This policy appears to have worked and by spring 1999, Bulletins were again appearing on time.

Between May 1993 and March 2000 there have been 23 article submissions, averaging at about 3 papers per year over 7 years. These articles have varied in length from 4 to 45 Bulletin pages, with the average length of articles being 20 Bulletin pages. On this basis, each Bulletin should contain approximately 3 x 20 pages, i.e. 60 pages.

To give these figures some context, it is appropriate to compare them with data from the previous ten years (1983-1993; Bulletins 33-41, but not including the Index in Bulletin 36). During this time, 41 papers were published (about 4 submissions per year), with an average length of 18.5 Bulletin pages. This resulted in Bulletins 33-41 having an average length of around 80 pages.

The two data sets are not directly comparable because today we typically print more words per page. However, the general trend suggests that submissions are down a little, although today the papers are, on average, a bit longer. On balance these figures suggest that over the coming years, if the rate of article submission remains at present levels, the society should be able to publish yearly Bulletins with about 60 pages.

As we move into the new millennium I sincerely hope that the rate of submission, and the high quality of submissions is sustained, allowing the Society to continue to produce its distinctive journal. I therefore welcome the continued submission of papers on all aspects of East Anglian geology.

Thank you to reviewers

This is an appropriate time to formally thank article reviewers who have given freely of their time over the last eight years to ensure that the Bulletin maintains an appropriately high scientific quality. The following reviewers have given excellent advice, many on more than one occasion.

Dr. T. Atkinson, Dr I. Boomer, Prof. P. Doyle, Prof. B. Funnell, Dr. R. Gallois, Dr. R. Gehrels, Dr. P. Gibbard, Prof. J. Hancock, Dr. C. Harris, Dr. P. Hoare, Prof. M. Leeder, Dr. D. Martill, Prof. J. Orford, Prof. R. Raiswell, Prof. I. Shennan, Prof. R. West.

I sincerely apologize if anyone has been accidentally missed from this list: I value all reviewing contributions.

Acknowledgments

The publication of Bulletin 50 is also an appropriate time to thank the School of Environmental Sciences at the University of East Anglia for its contribution to the production of the Bulletin. Three Deans of the School (Prof. Keith Clayton, Prof. Fred Vine and Prof. Trevor Davies) are thanked for their kindness in providing free secretarial and reprographic support.

I am indebted to Prue Barton, Fiona Woodgate, Julie parsons, Chris Flack and Rosie Cullington who have all helped at various times with word processing. Sheila Davies has assisted with half tone photographs and Philip Judge has, on a number of occasions, transformed line drawings.

The late Barbara Slade is especially remembered and thanked for her years of hard work on the Bulletin in the 1980s when Prof. Brian Funnell was editor.

INSTRUCTIONS TO AUTHORS

If possible, contributors should submit manuscripts as word-processor print out accompanied by a disk copy. We can handle most word-processing formats although PC Word, WordPerfect or ASCII files are preferred. In addition we accept typewritten copy and will consider legible handwritten material.

It is important that the style of the paper, in terms of overall format, capitalisation, punctuation, etc. conforms as strictly as possible to that used in Vol. 50 of the Bulletin. Titles and first order headings should be capitalised, centred and in bold print. Second order headings should be centred, bold and lower case. Text should be 1½ line spaced. All measurements should be given in metric units.

References should be arranged alphabetically in the following style.

BALSON, P.S. & CAMERON, T.T.J. 1985. Quaternary mapping offshore East Anglia. *Modern Geology*, **9**, 221-239.

STEERS, J.A. 1960. Physiography and evolution: the physiography and evolution of Scolt Head Island. In: Steers, J.D. (ed.) *Scolt Head Island* (2nd ed.), 12-66, Heffer, Cambridge.

BLACK, R.M. 1988. *The Elements of Palaeontology*. 2nd Ed., Cambridge University Press, Cambridge. 404pp.

Illustrations should be drawn with thin dense black ink lines. Thick lines, close stipple or patches of solid black or grey should be avoided as these can spread in printing. Original illustrations should, before reproduction, be not more than **175mm by 255mm**. Full use should be made of the first (horizontal) dimension which corresponds to the width of print on the page, but the second (vertical) dimension is an upper limit only. Half tone photographic plates are acceptable when their use is warranted by the subject matter, provided the originals exhibit good contrast.

The editors welcome original research papers, notes or comments, and review articles relevant to the geology of **East Anglia** as a whole, and do not restrict consideration to articles covering Norfolk alone. All papers are independently refereed by at least one reviewer.

C. E. RANSON'S DATA FROM THE GLACIFLUVIAL AND OTHER SANDS AND GRAVELS OF NORTH NORFOLK, ENGLAND

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ABSTRACT

General observations and clast lithological data collected during the 1970s by the late Colin Ranson suggest that the widespread sands and gravels of north Norfolk were laid down during the melting back of the (second?) Lowestoft ice sheet by currents flowing broadly to the south-east.

INTRODUCTION

The untimely death of Colin Ranson (1936-1989) came before he could complete and publish the results of his already extensive survey of the sands and gravels which stratigraphically overlie the Cromer Tills (North Sea Drift, Norwich Brickearth) of north Norfolk. Ranson assiduously collected these data while on family holidays and as he travelled all over East Anglia as a locally-based Assistant Regional Officer of the then Nature Conservancy Council. This work, largely on inland sites, was itself a development from an earlier period of study of the Cromer tills and intervening waterlaid beds of the Norfolk coastal cliffs (e.g. Ranson, 1968; Banham & Ranson, 1965; 1970). Incidentally, as they travelled around East Anglia the Ransons also noted sites of interest for landscape history; Oliver Rackham (1986) recognised the importance of these observations when he dedicated his *'History of the Countryside'* to "Colin and Susan Ranson, my trusty friends and helpers".

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No detailed interpretations were left by Ranson, and none is attempted here. However, these data from 38 sites have both an intrinsic value and an increased significance as many of the sand and gravel pits he investigated are no longer available to researchers (see e.g. Fig. 1). For that matter, the coastal cliff exposures are now much poorer than they were up to thirty-five years ago, before modern sea-defences were put in place. Ranson's work throws light upon the lateral extent, thickness, facies, palaeocurrent flow directions, and, in particular, the composition of the pebble fraction of these sands and gravels. Also presented here for completeness and for purposes of comparison, are the data he collected from seven sites in other, generally younger gravel deposits, including the Blakeney esker, Morston raised beach and two modern beaches.

Fig. 1. Anglian Sands and Gravels in the Briton's Lane Pit, near Sheringham (1964; PHB) A view to the south from approximatley TG167415 (see Appendix 1, localities 14 and 15).

THE ANGLIAN GLACIFLUVIAL SANDS AND GRAVELS

Sediment descriptions and palaeocurrent indicators

For 20 of the 38 sites within these sands and gravels Ranson provided a detailed description covering bed thicknesses, grain sizes and, where possible, palaeocurrent indicators (Appendix 1). From these descriptions, this extensive unit of sand and gravels is known to have a thickness of at least 22.75m (Appendix 1, locality 10). In the eastern portion of their outcrop the sands and gravels rest upon sandy Cromer Till(s) while to the west, they both overlie and wedge into the chalky Lowestoft Till. In places, both this till and the sands and gravels have been folded (e.g. Appendix 1, localities 31, 33 & 38). These sands and gravels thus appear to have been laid down after the deposition of the Cromer Tills and penecontemporaneously with the Lowestoft Till.

These data also show that grain size generally increases to the north and west (Fig. 2). Consistent with that, measurements of fifty-five flow direction indicators from nine sites generally indicate palaeocurrent flow to the south-east (Figs. 2 & 3).

Clast Lithological Analyses

'Pebble counts' were the main thrust of Ranson's work. A total of 13,757 determinations (8-32 mm, intermediate dimension) of pebble composition is given for the 38 sample sites within the Anglian glacifluvial sand and gravel unit (Table 1). The dominant component by far is flint, which reaches its highest values in the west (92- 99%, Fig. 4), where the sands and gravels directly overlie flint- (and chalk-) rich Lowestoft Till and flint-rich Upper Chalk. To the east, flint is somewhat less dominant (81-92%, Fig. 4), with a compensatory increase in quartz and quartzite, the components next in importance (Fig. 5a). Here the sands and gravels overlie relatively quartz and quartzite-rich Cromer Tills which in turn overlie similarly quartz and quartzite-rich 'pre-glacial' sands and gravels (e.g. Briant *et al.*, 1999; Rose et al., 1999; Green & McGregor, 1999). Pronounced chatter-marking of flint pebbles found at only seven sites also appears to increase from east (2%) to west (7%; cf. modern raised beaches which have up to 72% chatter-marked flints, Table 1).

No other clear linear relationships have been found between altitude and lithology (Fig. 5b) nor between lithology and lithology; this remains true even when the data are

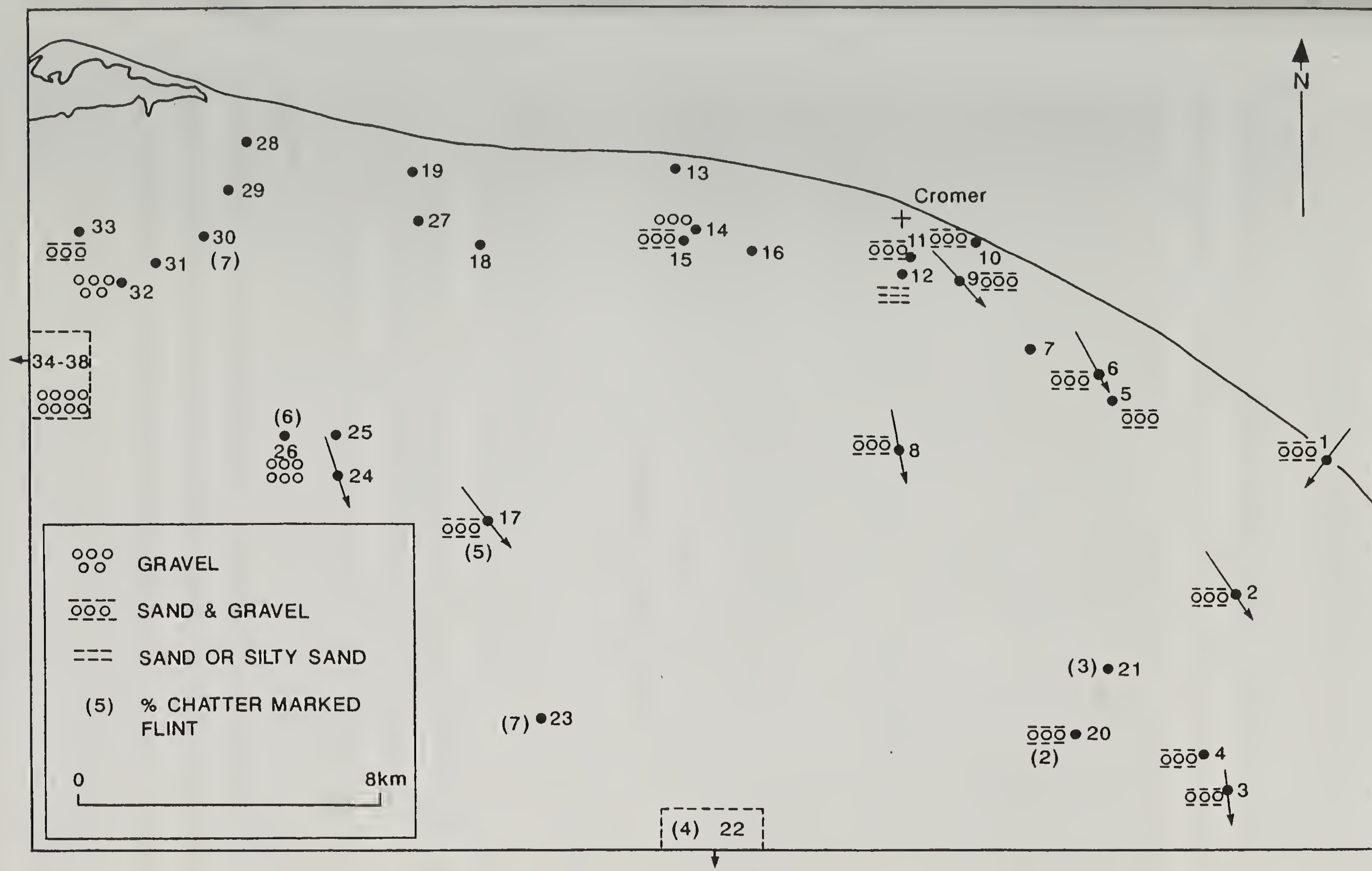


Fig. 2. Sample localities in the Anglian glacifluvial gravels of north Norfolk (Appendix 1 localities 1-38). Grid references, grain size and facies indications from Appendix 1. Numbers inset are for sample localities to the south (22) and west (34-38) of the map area. Numbers in brackets represent % chatter-marked flints (6 localities only). Arrows indicate probable/possible palaeocurrent flow directions (and see Fig. 3).

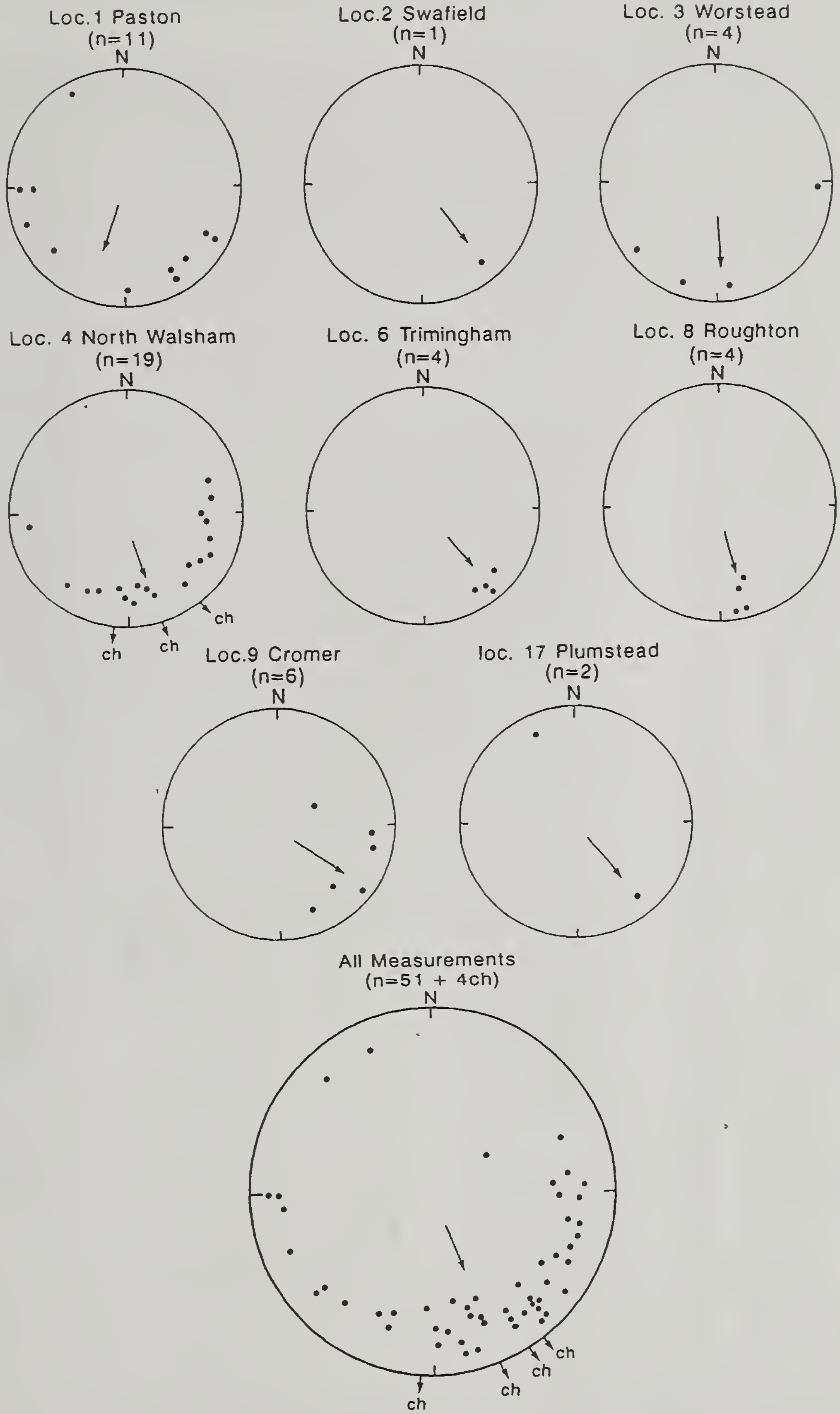


Fig. 3. Palaeocurrent indicators within the Anglian glacifluvial gravels of north Norfolk. Wulff projection, lower hemisphere; poles indicate direction and amount of dip of cross-bedding foresets (where no dip amount was recorded (Appendix 1), a value between 5-10 has been assumed). ch = fluvial channels; arrows indicate inferred current directions (see Fig. 2).

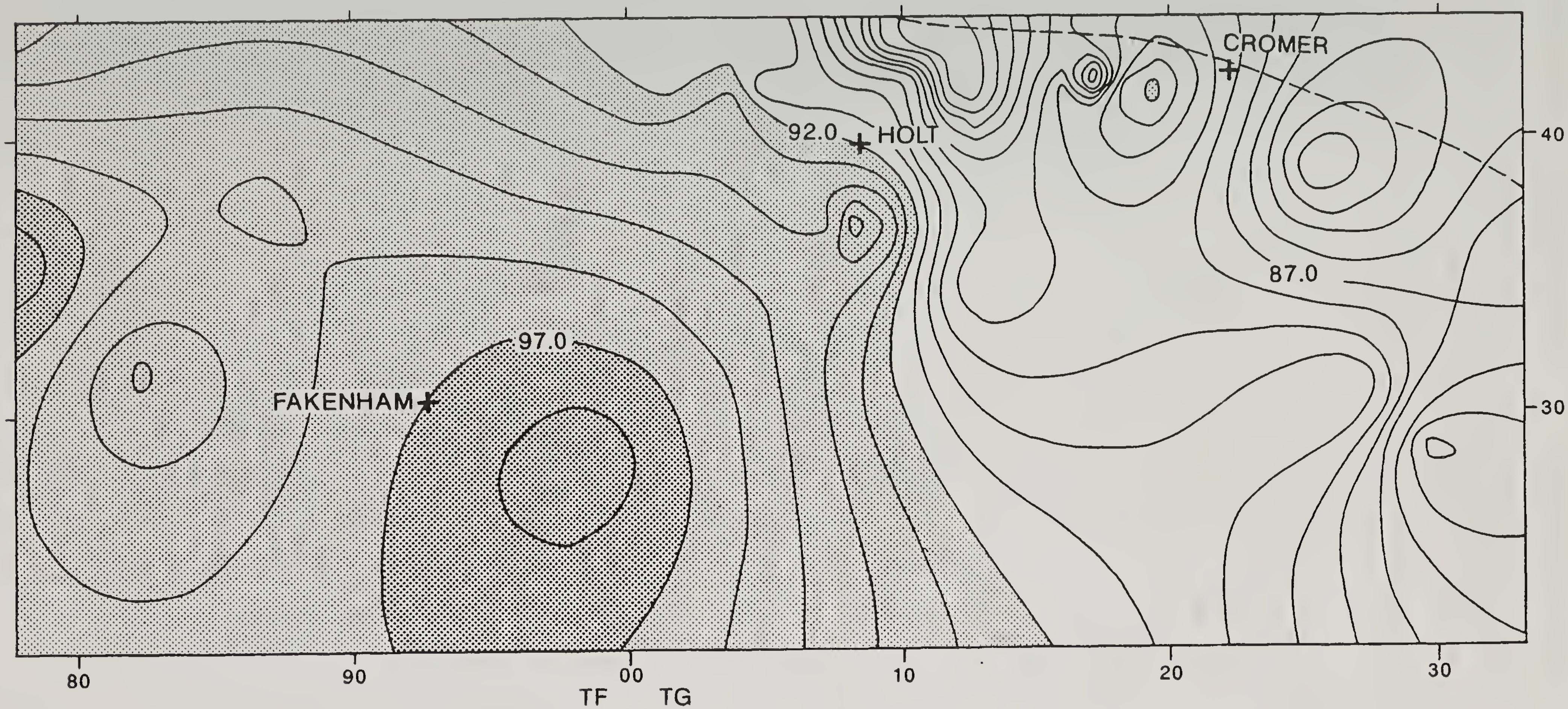


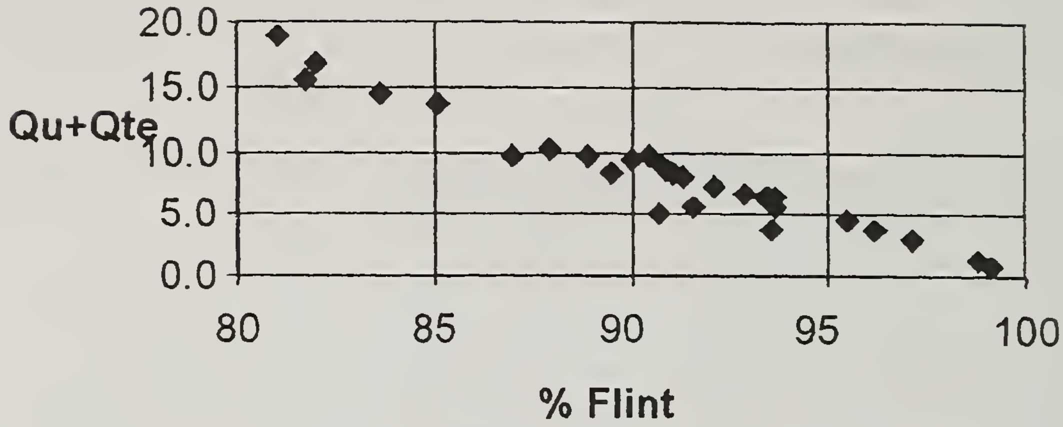
Fig. 4. Computer plot (Surfer©) of flint percentages in the Anglian glacifluvial gravels of north Norfolk. Coastline indicated by pecked line. The boundary between the Ordnance Survey TF and TG map areas is shown, along with national grid coordinates. The 92% isopleth approximately coincides with the boundary between substrates, i.e. chalk and flint rich to the west and relatively quartz and quartzite rich sandy deposits to the east.

Table 1: Ranson's clast analyses

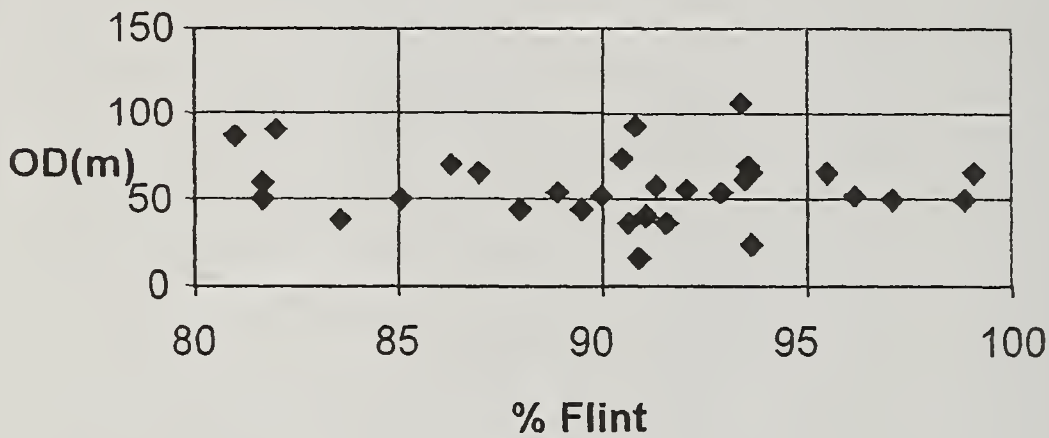
Locality No.	OD(m)	%Flint	%Quartz	%Quartzite	%Chalk	%Meta	%Other	Total Pebbles
1	43	88.0	6.4	3.7	0	0	0	266
4	38	83.6	3.8	10.6	1.0	0	0	104
7	49	81.7	7.7	7.7	0	0	1.0	169
8	49	85.1	9.1	4.6	0.6	0	0.6	154
10	69	86.3	nr	nr	nr	nr	nr	327
13	54	88.9	5.9	3.7	0.4	0	1.1	270
14	90	82.0	5.1	11.6	0.6	0.8	0	511
15	92	90.8	2.4	6.5	0.2	0	0	382
16	107	93.4	2.5	3.8	0.3	0	0	319
17	66	87(5)	3.9	5.7	2.0	0.2	1.2	507
18	87	81.0	7.5	11.5	0	0	0	295
19	59	81.7	7.7	7.7	0	0	1.0	846
20	36	90.7(2)	3.8	1.3	0.5	0.2	0	450
21	39	91.1(3)	4.0	4.2	0.4	0.2	0	1039
22	36	91.6(4)	2.4	3.2	1.3	1.5	0	532
23	51	90.0(7)	3.6	5.7	0	0.2	0.6	611
24	66	95.5	0.4	4.1	0	0	0	269
25	49	97.1	1.1	1.8	0	0	0	278
26	69	93.6(6)	2.4	1.4	2.4	0.1	0.1	675
27	74	90.5	1.5	8.0	0	0	0	336
28	16	90.9	3.3	5.3	0	0	0.5	430
29	56	92.1	3.2	4.1	0	0.6	0	317
30	43	89.5(7)	3.0	5.2	1.9	0	0.4	687
31	23	93.7	0.8	4.9	0	0	0	366
32	61	93.5	1.5	4.9	0	0	0	398
33	57	91.3	3.4	4.6	0	0.5	0	391
34	49	98.8	0.4	0.8	0	0	0	728
35	51	96.2	1.3	2.5	0	0	0	768
36	66	93.7	0	6.3	0	0	0	254
37	66	99.1	0.2	0.7	0	0	0	696
38	53	92.9	1.6	5.0	0.5	0	0	382
Total								**13757**
39	46	97.5(12)	1.1	1.4	0	0	0	823
40	33	95.6(15)	2	2.3	0	0	0	856
41	13	87.2	4.2	6.1	0	2.5	0	475
42	13	83.0(8)	7.9	7.3	0.5	0.5	0.8	848
43	5	99.2(72)	0.4	0.4	0	0	0	253
44	0	95.0(68)	1.3	3.3	0	0	0.4	758
45	0	84.0(27)	6.8	5.6	0.6	0	0	161
Total								**17931**

Clasts in the size range 8-32 mm, intermediate axis. For grid references and descriptions of (some) sites, see Appendix 1. Meta = metamorphics; nr = not recorded/not available; values in brackets e.g. (7) = % chattermarked flints.

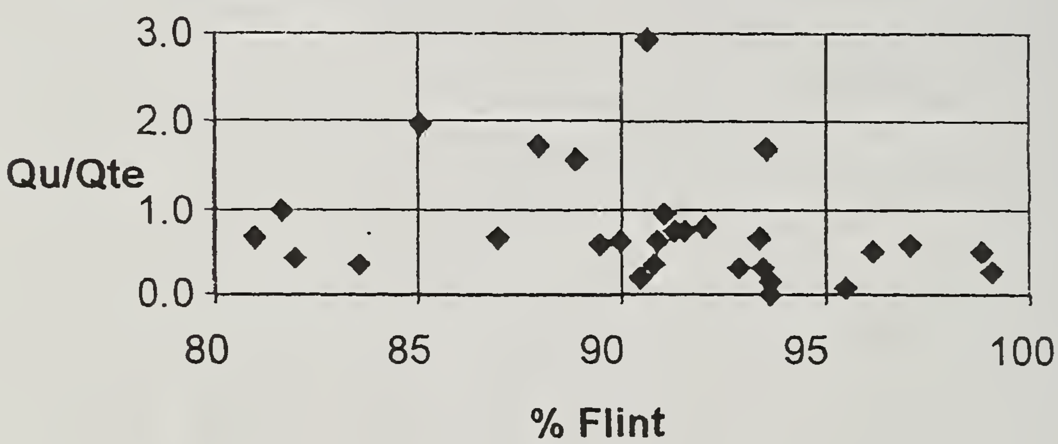
Locality nos. 1-38 are Anglian glacifluvial gravels; locality nos. 39, 40 are Blakeney esker gravels; locality nos.41, 42 are 'Valley Gravels (post -Anglian); locality no. 43 is Morston raised beach (?Ipswichian); modern beaches are represented by locality nos. 44 (Weybourne) and 45 (Mundesley).



a. Quart plus quartzite/flint: the clear linear relationship found is to be expected between the two major components.



b. Altitude/flint : there is apparently no simple relationship.



c. Quartz : quartzite ratio/flint: this plot gives further support for more mixed origins for the relatively flint-poor eastern province (see text and Fig. 4).

Fig. 5. Clast lithology in the Anglian glacifluvial gravels of north Norfolk. (Note: when these and other variables are plotted separately for eastern and western provinces, no further simple relationships are revealed).

plotted separately for the eastern and western provinces, suggested when the quartz to quartzite ratio is plotted against flint (Fig. 5c).

Interpretation

From these data a necessarily speculative and general interpretation of the environment of deposition of this sand and gravel unit seems to require the melting back of an ice sheet to the west and north and the deposition of a sandur from meltwaters flowing broadly to the south-east (Fig. 6). It is noteworthy that the headwaters of the modern rivers Bure and Ant still drain this area to the south-east. The involvement of the Lowestoft ice seems most likely, perhaps the second Lowestoft advance of Ehlers et al.(1987). Continued melting of the Cromer ice to the north (Banham, 1988) is also possible; conceivably, lobes of these two ice sheets had merged before this time. More speculative still is the suggestion (Fig. 6) that the several, approximately N-S linear hills underlain by down-folded sand and gravel-filled 'basins' within the 'hummocky drift' area of coastal north

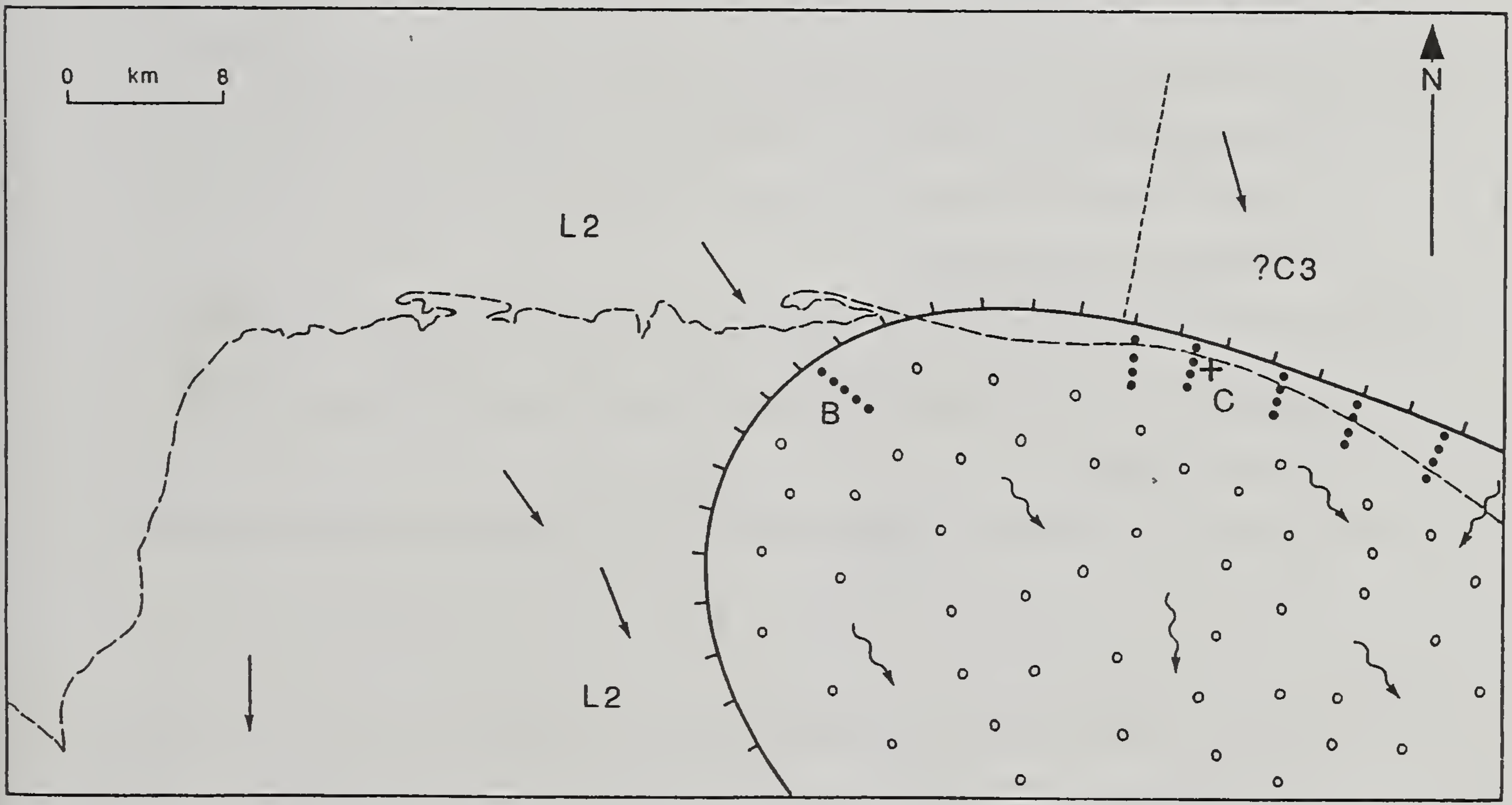


Fig. 6. Speculative reconstruction of the depositional context of the Anglian glacifluvial gravels of north Norfolk. Pecked line = modern coastline; C = Cromer; wavy arrows = palaeocurrent flow directions; straight arrows = ice movement directions; L2 = second advance, Lowestoft ice (Ehlers et al.,1987); ?C 3 = possible third advance, Cromer ice (Banham,1988); B = Blakeney esker. N-S dot symbols = N-S linear hills (see text).

Norfolk (e.g. Banham,1975) reflect the deposition and loading by ice-marginal esker streams feeding a glacifluvial outwash plain to the south. Near Norwich, Funnell (1976) has proposed a similar relationship between a buried valley, the Mousehold Heath outwash plain and a Lowestoft ice margin. Further, on the basis of its position, orientation and broadly consistent pebble compositions at only two localities (Appendix 1 localities 39 & 40; Tables 1), it is possible that the well-known Blakeney esker could also be of this origin (Fig. 6), as suggested by Rose in Banham *et al.* (1975).

Colin Ranson is not responsible for these inadequately supported interpretations; I can almost hear him saying, ‘Clearly, much more work is needed here’.

ACKNOWLEDGEMENTS

The author is grateful for considerable assistance with the application of computer technology, especially in the preparation of the figures: Clare Banham for Figs. 1, 2, 3, 4, and 6; Janet Banham for Figs. 4, 5a-c and Table 1.

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APPENDIX 1. DESCRIPTION OF SITES

A transcript of Ranson's 'Details of Exposures', plus locations and brief notes on sites for which only pebble count (i. e. clast lithology) data is available. Sites are thought to be within Anglian glacifluvial deposits (1-38) unless otherwise stated (39-45). All dip directions are in degrees.

Locality 1. Paston TG 323359 Coastal cliff section.

13.0 m	yellow cross-stratified sand with occasional pebble trains, especially in the lower 2m; sets 100-300 mm thick
0.6 m	gravel
1.1 m	cross-stratified sand
0.2 m	ferruginous gravel

Cross-strata dip directions to 150, 120, 140, 270, 270, 150, 120, 180, 330, 230, 250.

Locality 2. Swafield TG 298330 Old gravel pit, now filled in.

0.70m cross-stratified yellow sands with a few stones; units 50-80mm thick

Cross-strata dip directions to 10 to 140.

Locality 3. Worstead, Sandy Hill TG 296280 Old gravel pit.

An almost identical succession to that at Scarborough Hill (Locality 4).

Locality 4. North Walsham, Scarborough Hill TG 290288 Sand and gravel pit.

6 m + cross-stratified and laminated yellow sands with many pebble trains and gravelly beds - especially near the top. The sets of cross-strata are never more than 300-400 mm thick. A few silty, ferruginous horizons up to 200 mm thick are present. A number of channels cut through the cross-stratified sands; they are filled with gravelly sand.

Cross-strata maximum dip directions 16 to 85, 15 to 105, 22 to 90, 15 to 70, 20 to 95, 18 to 205, 11 to 115, 15 to 120, 18 to 125, 15 to 220, 12 to 140, 16 to 160, 20 to 200, 17 to 180, 12 to 265, 15 to 175, 27 to 170, 25 to 185, 23 to 165.

Channel directions to 145, 185, 160.

Locality 5. Southrepps, Ash Tree Farm TG 268377 Roadside exposure.

1.0 m loam and soil

0.4 m sand and pebbles

Locality 6. Trimingham TG 264383 Old railway cutting.

2.85+ m cross-stratified yellow-orange sands with frequent pebble beds 50-120 mm thick . Units are less than 150 mm thick.

Cross-strata maximum dip directions 5 to 140, 12 to 130, 10 to 145, 5 to 140.

Locality 7. Northrepps TG 250387

Locality 8. Roughton, A140 roadside exposure at TG 218365 and, about 6 m below, a temporary exposure opposite the New Inn at TG 220369.

0.6 m former road footings

2.0 m cross-stratified yellow sands with ferruginous horizons 20-40 mm thick. Pebble beds , especially near base, up to 50 mm thick. Units 300-400 thick.

0.15 m silty sand with carbonaceous patches.

0.65 m laminated yellow-orange sand; laminae about 10 mm thick.

0.20 m silty brown sand with thin (20 mm) clay seam

0.60 m laminated yellow sands with a 20 mm thick pebble bed; laminae about 10 mm thick

Cross-strata not seen in the roadside exposure, but a dip of 8 to 80 was measured. Cross-strata in the New Inn temporary exposure, maximum dip directions 25 to 160, 5 to 165, 5 to 170, 20 to 165.

Locality 9. Cromer, Northrepps Road. TG 231407 Old gravel pit, 40m wide.

West side of pit

2.0 m undisturbed cross-stratified sands with four pebble beds 100-200 m thick, and a ferruginous silty bed 20 m thick

Cross-strata maximum dip directions 50 to 60, 12 to 95, 5 to 130, 15 to 160, 20 to 140, 10 to 105.

East side of pit

4.0 m contorted sands with nests of boulders

Locality 10. Overstrand, Kirby Hill TG 235415 Coastal cliff section.

6.75 m coarse orange gravel, waterlain, well-bedded, but showing no cross-stratification. Distinct pebble orientation. Sharp boundary with underlying unit

16.0 m pale sands and gravels. These are made up of 11 m of sands, sometimes cross-stratified, in beds 150-500 mm thick, and 5 m of gravels in beds up to 400 mm thick. The base is gravelly and a proportion of the pebbles are chalk. Shell patches?

Locality 11. Cromer, Hillside Road. TG 22241 Temporary exposure on building site.

0.65 m hard pebbly silty sand

1.30 +m laminated yellow sand with no pebbles

Locality 12. Cromer, Burnt Hills TG 219407 Temporary exposure on building site.

1.5+ m	yellow-brown fine silty sands. These silty sands are very compact and form steep banks. A similar bed 1+m thick is seen 100 m to the east. Loess?

Locality 13. Sheringham TG 168434

Locality 14. Briton's Lane Gravel Pit, *north end* TG171417

20 m	flat-bedded gravels dipping 2- 5 to 0

Locality 15. Beeston Regis, Briton's Lane TG 167415. Sand and gravel pit.

East side

8.0 m	sandy gravel passing up into coarse gravel
0.1 m	grey clay
2.0 m	shelly sand and gravel
5.0 m	sands and gravels

South side

4.0 m	gravel and sand
1.0 m	chalky clay (till)
3.0+m	yellow sands with pebble beds.

Locality 16. Aylmerton TG 184415

Locality 17. Plumstead, The Heath TG 122346

3.0+ m	well-bedded, cross-stratified gravelly sands.

Cross-strata dip toward 140, 340.

Locality 18. Weybourne Heath TG 121414

Locality 19. Muckleburgh Hill TG 100428

Locality 20. Felmingham, Bryant's Heath TG 257295

1.70+m	orange bedded sands with pebble trains.

Locality 21. North Walsham TG 266310

Locality 22. Buxton Heath TG 173217

Locality 23. Oulton TG 135297

Locality 24. Edgefield TG 086357 Sand and gravel pit.
Flow direction to 150 from imbrication and channels

Locality 25. Edgefield Hill TG 087367

Locality 26. Edgefield Heath TG 074366

Locality 27. Kelling Heath TG 102420

Locality 28. Salthouse TG 062440

Locality 29. Cley TG 058426

Locality 30. Glandford TG 054416

Locality 31. Glandford TG 043411 Sand and gravel pit.
Fold with core of chalky boulder clay; axis to 155.

Locality 32. Saxlingham TG 033404. Sand and gravel pit (?)
Coarse gravels with lenses of chalky boulder clay; 35 dip to 195.

Locality 33. Langham, Bilsey Hill TG 021417 Sand and gravel pit.
Flat-bedded coarse ferruginous sands with lenses of chalky boulder clay; 10 dip to 130 fold axis to 170.

Locality 34. Hempton Green TF 911286

Locality 35. North Creake TF 877381

Locality 36. Syderstone Common TF 833317

Locality 37. Docking Common TF 791355

Locality 38. Brancaster, Barrow Common TF 792431 Sand and gravel pit.
Bedded gravels overlying chalky boulder clay, fold - axis 135, overturned to SW.

Locality 39. Wiveton Downs TG 030423 Sand and gravel pit.
Blakeney Esker gravels.

Locality 40. Morston TG 019438 Old pit.
Blakeney Esker gravels.

Locality 41. Weybourne Hope TG 112437 Coastal cliff top.
(Devensian?)'Valley Gravels'.

Locality 42. West Runton TG 186433 Coastal cliff top.
(Devensian?)'Valley Gravels'.

Locality 43. Morston TF 991441
(Ipswichian?) raised beach.

Locality 44. Weybourne TG 110433
Modern beach.

Locality 45. Mundesley TG 323360
Modern beach.

[Manuscript received 21 February 2000; Revision accepted 23 February 2000]

THE GLACIAL GEOLOGY OF THE WEYBOURNE AREA, NORTH NORFOLK: A NEW APPROACH

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ABSTRACT

Study of till sections in the Weybourne area of north Norfolk revealed a succession of brown and pale yellow laminated diamictons. In addition to traditional sedimentary analyses, a new technique, based on chalk micropalaeontology was employed to provenance both chalk clasts and chalk microfossils from the matrix in order to aid interpretation of sediments. The results show that chalk material from Weybourne till has been sourced primarily from the west but that some had an eastern source. The sediments at Weybourne are interpreted as being the result of a two-fold glacial advance. The first advance, by Scandinavian Ice associated with the Happisburgh Diamicton, deposited a till primarily composed of reworked crag with some intermixed eastern material. The second advance from the south-west, by the British Ice Sheet, deformed and glaciotectonised the Crag-rich till, mixing in western erratics and resulting in the complex sedimentary sequence seen at Weybourne. All the tills sampled are therefore correlated with the Lowestoft Till.

INTRODUCTION

It is well known that north-east Norfolk possesses spectacular cliffs which have interested glacial geologists for many years (Reid, 1882; Solomon, 1932; Dhonau and Dhonau, 1963; Banham and Ranson, 1965; Hart and Boulton, 1991a; Lunkka, 1994), and yet there are still significant differences of opinion regarding the origin of their sediments and the sequence of glacial events which they reveal (Eyles et al., 1989; Hart and Boulton, 1991a; Rose, 1992; Lunkka, 1994). Glacial sediments exposed in Weybourne Town Pit (TG 114 431) and in coastal cliffs immediately to the north (TG 108 438 & TG 113 437) are of particular note. They show inter-bedding of distinctly different diamictons and this

Bull. geol. Soc. Norfolk (for 2000) **50**, 21-45 (Published 2000)

complex structural and stratigraphic relationship has been used to infer that the area around Weybourne may straddle former coalescing and fluctuating margins of British and Scandinavian ice sheets (Hart and Boulton, 1991a, figure 158e). Glaciotectonic structures in the Weybourne cliffs have been used to infer local ice flow directions (Banham, 1965, Banham and Ranson, 1965), but no detailed provenance work has been carried out on the tills of the Weybourne area and some details of its glaciological development remain unresolved.

Recently, a new approach to till provenance has been examined by members of the School of the Environment at the University of Brighton in collaboration with the British Geological Survey (BGS). This has thrown new light on the problem of ice sheet movement, in north Norfolk in particular, and throughout Eastern England (Moorlock et al., in press). The new technique involves the use of chalk micropalaeontology to provenance chalk erratics from till units in relation to the outcrop of Chalk foraminifera biozones (Fig. 1). In an earlier attempt to provenance chalk erratics Peake and Hancock (1961) used macrofossils, but the present work is the first time that micropalaeontology has been applied in the investigation of chalk in British tills. This paper reports the results of an investigation using the new technique, supplemented by more conventional analyses. Its aim is to determine the provenance of tills, the direction of ice flow and the relative contribution of British and Scandinavian ice sheets to the deposition of till around Weybourne, north Norfolk.

METHODOLOGY

The main source of data for this investigation is Weybourne Town Pit. The section was cleaned, drawn and logged to record the sedimentary sequence (Fig. 2). Samples were obtained from each sedimentary unit for particle size analysis, calcium carbonate determination and clast lithological analysis (Tables 1 and 2). Colours were recorded from fresh exposures using Munsell notation. Clast macrofabric analysis was conducted on clasts over 1 cm in length, with a:b axis ratios of at least 3:2. The strike direction of prominent fold noses and the poles to planes of fold limbs were recorded to assess the direction of deformational stress. Chalk clasts, where present, and samples of diamicton matrix were obtained for micropalaeontological provenance analysis. Additional micropalaeontological material was collected from diamictons in the coastal cliffs 0.8 km north of Weybourne village, in order to assess the spatial consistency of the Town Pit evidence.

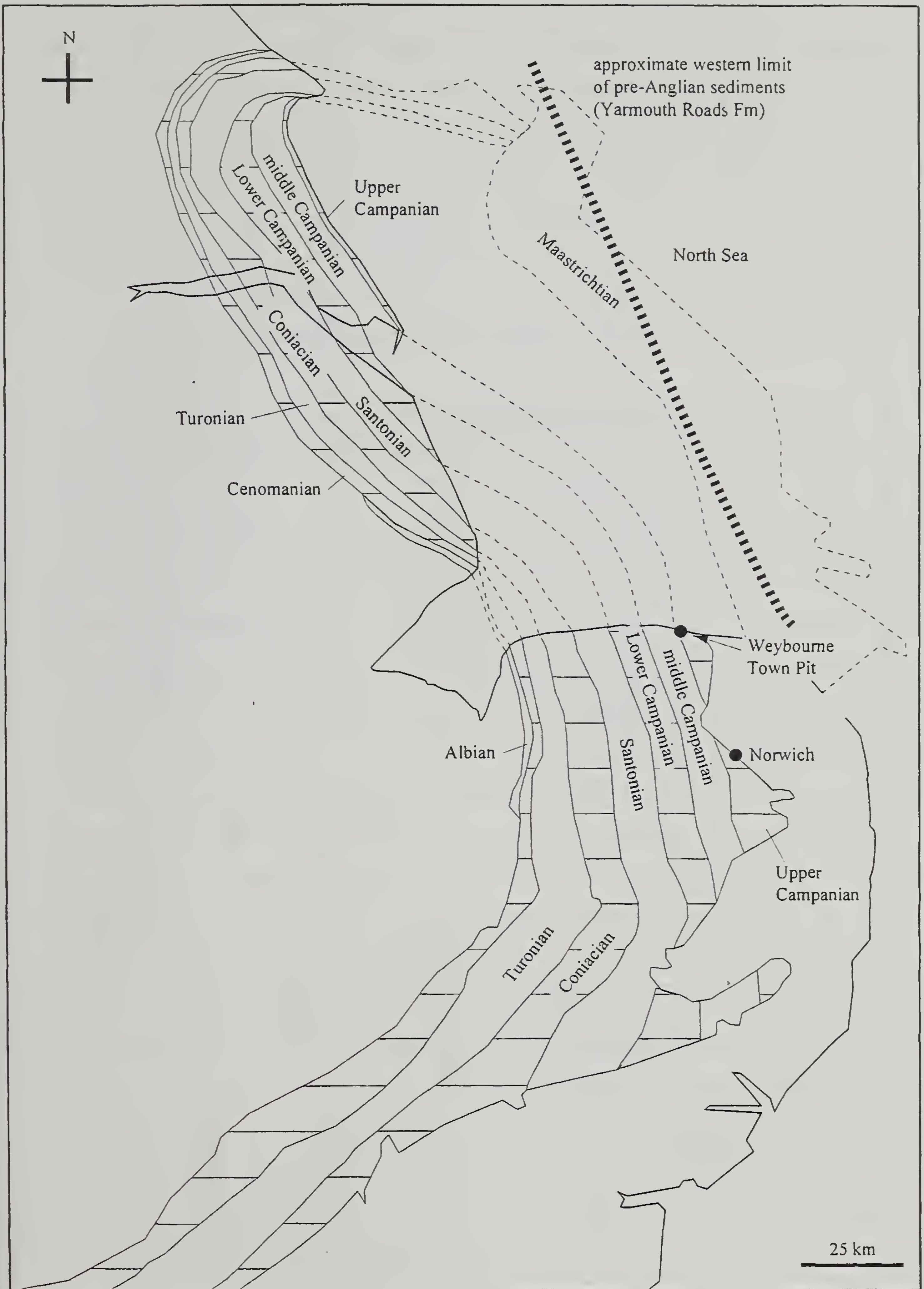
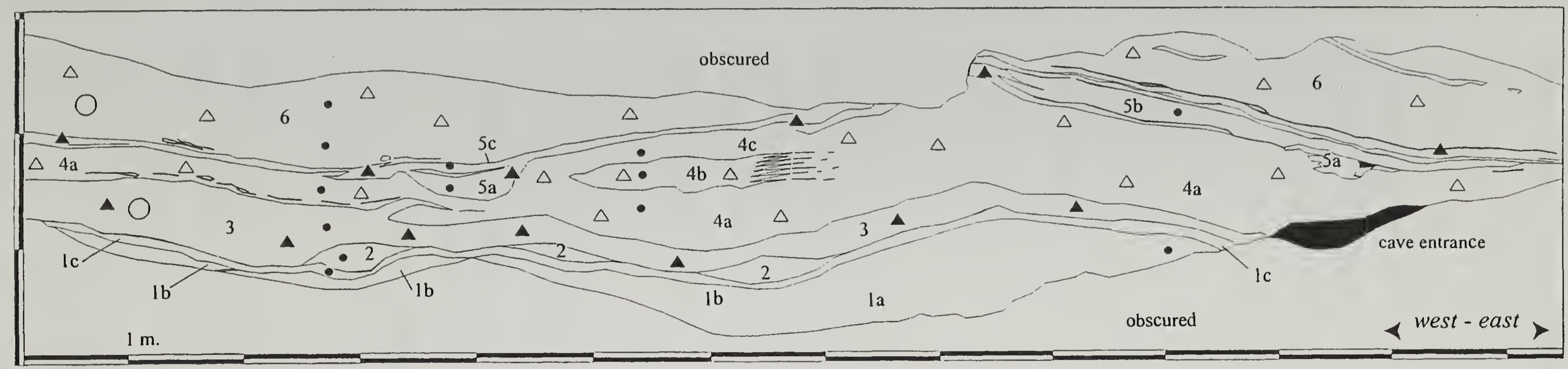


Fig. 1. Outcrop/subcrop of Chalk zones in eastern England and adjacent North Sea.



Unit	Description
6	Silty till (upper), pale yellow (2.5Y7/4)
5c	Sandy till (upper), brown (10YR 4/3) to yellowish brown (10YR 5/4) and dark yellowish brown (10YR 4/4)
5b	Chalky lens, very pale brown (10YR 8/2)
5a	Sandy till (upper), yellowish brown (10YR 5/4)
4c	Silty till (lower), pale yellow (2.5Y 7/4)
4b	Silty till, (lower), very pale brown (10YR 7/3)
4a	Silty till, (lower), pale yellow (2.5Y 7/4, 2.5Y7/3)
3	Sandy till (lower), yellowish brown (10YR 5/8, 10YR 6/8)
2	Silty sand, yellow (10YR 7/8) to pale yellow (2.5Y 7/4)
1c	Chalk marl (upper), very pale brown (10YR 8/2)
1b	Chalk marl (lower), very pale brown (10YR 8/2) to pale yellow (2.5Y 7/3)
1a	Banded chalk marl, pale yellow (2.5Y 8/2) with very pale brown (10YR 7/4 laminations)
•	Sample location for particle size and clast lithological analyses
○	Clast macrofabric location
▲	Brown till
△	Pale yellow till
△	Composite till

Fig. 2. Section logged at Weybourne Town Pit, June 1998. Note that the marginal black and white scale bars represent 1 m.

Samples for micropalaeontological analysis were prepared using the technique employed by BGS personnel at Keyworth (Ian Wilkinson, personal communication). Fossils were identified using a top-lit binocular microscope with magnification of up to x80. Fossil identification was aided by photographs and range charts from the literature (Koch, 1977; Bailey, 1978; Robasyznski et al., 1983; Swiecicki, 1980; Hart et al., 1989) and a reference collection from the BGS. The technique used to provenance fossils obtained from both chalk clasts and chalky matrix relies on three concepts:

1. that the chalk biozones outcrop at known locations (Fig. 1).
2. that these biozones can be identified by a characteristic foraminifera assemblage.
3. that the age of individual chalk samples can be constrained by reference to the stratigraphical age ranges of fossils.

A brief example will illustrate the method of ageing a sample (Fig. 3). A number of foraminifera species are identified and their stratigraphical ranges determined. *Globorotilites micheliniana* is of no stratigraphical value as it is too long-ranging. The other species are of more use because of their shorter stratigraphical ranges. *Stensioeina exculpta gracilis* indicates that the age of the chalk sample must be no earlier than the early *Uintacrinus socialis* zone of the Santonian Stage. However, the presence of *Stensioeina pommerana* indicates that the sample cannot be from the *U. socialis* biozone, or an earlier biozone as *S. pommerana* first appears in the middle of the *Offaster pilula* zone, in the early Campanian Stage. Therefore the earliest age, constrained by the latest species appearance, is equivalent to the upper *O. pilula* biozone. To constrain the upper age limit of the chalk sample *G. micheliniana* and *S. exculpta gracilis* are of little use. However, the disappearance of *Bolivinoides strigillatus* at the top of the *O. pilula* biozone indicates that the sample must be from the Lower Campanian and by using the appearance of *S. pommerana,* an even more accurate dating, of upper *O. pilula*, can be obtained, as shown by the shaded box in Figure 3. The age of the chalk clast is therefore upper Lower Campanian and its provenance is within the outcrop of this biozone. This technique is analogous to the mutual climatic range method for comparing sediments using beetle remains (Lowe and Walker, 1997).

In this study the Campanian Stage of the Upper Cretaceous has been subdivided differently to that usually found in the literature. Here, it is subdivided into upper, middle

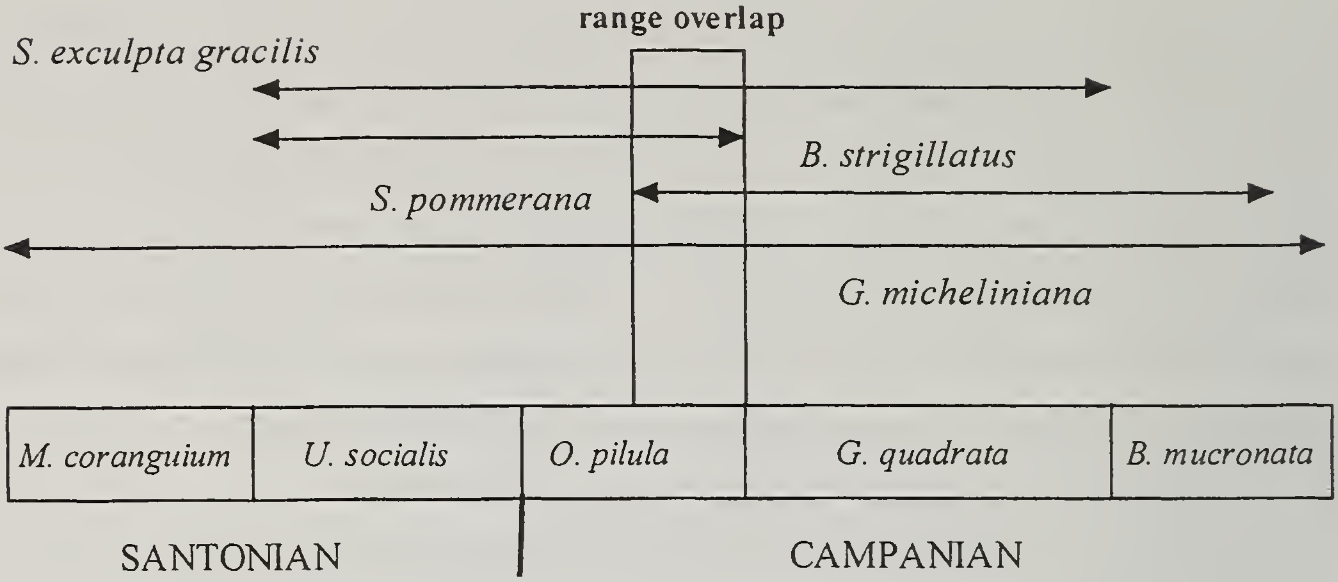


Fig. 3. Method of derivation of Chalk age from a foraminiferal assemblage.

and lower substages, rather than the more usual Upper and Lower substages. The three substages used in this study correlate with to the biozones of *Offaster pilula* (Lower Campanian), *Gonioteuthis quadrata* (Middle Campanian), *Belemnitella mucronata* (Upper Campanian) (Bristow et al., 1997), which correlate with benthonic foraminifera biozones UKB 16, UKB 17 and UKB 18 (Hart et al. 1989).

SEDIMENTS OF WEYBOURNE TOWN PIT

The section at Weybourne Town Pit is illustrated in Figure 2 and sedimentological details are given in Tables 1 and 2 .

At the base of the exposure are two sorted sediments. Unit 1 is a clast-free, chalk-rich marl, up to 0.8 m thick, which can be subdivided by texture, structure and colour into three subunits, 1a, 1b, and 1c. The lowest subunit, 1a, is a laminated, pale yellow (2.5Y8/2) and very pale brown (10YR 7/4) clayey silt with very fine sand partings. The laminations are wavy and roughly parallel to the upper boundary of the unit. Subunit 1b has approximately equal parts of silt and clay and is very pale brown (10YR 8/2) to pale yellow (2.5Y7/3 in colour. The upper subunit, 1c, is a very pale brown (10YR 8/2) silty clay. Unit 2 is a yellow (10YR 7/8) to pale yellow (2.5Y7/4) calcareous sandy silt, forming discontinuous lenses or a very thin layer on top of unit 1. Its upper boundary is sharp and erosional. Together these sorted sediments are interpreted as waterlain, probably glaciofluvial sediment. The wavy boundaries and the stiffness of these units are attributed to gentle folding and compaction by glacial stress. The sharp upper contact and the discontinuity of unit 2 suggest that they have been partially eroded.

Table 1: Lithofacies, particle size and calcium carbonate content of sediments at Weybourne Town Pit.

Unit	*Lithofacies*	*Gravel Wt%*	*Sand Wt%*	*Silt Wt%*	*Clay Wt%*	*Modal Phi*	*$CaCO_3$ Wt%*
6	Pale till	12.78	29.07	33.93	24.22	8	62.70
5c	Brown till	1.12	37.90	46.07	14.91	4	11.07
5b	Chalk pod	0.00	0.42	62.70	36.88	8	82.72
5a	Brown till	26.74	42.68	24.40	6.18	4	14.65
4c	Pale till	4.88	25.84	42.43	26.86	8	67.25
4b	Dark pale till	5.75	32.89	35.90	25.46	3	55.79
4a	Pale till	8.48	28.81	40.59	22.13	8	63.36
3	Brown till	3.65	66.37	19.12	10.87	3	13.55
2	Silty sand pod	0.00	60.94	31.87	7.19	4	16.54
1c	Dark marl	0.00	2.21	29.82	67.97	9	67.54
1b	Light marl	0.00	1.84	48.00	50.16	8	73.41
1a	Banded marl	0.00	0.58	60.35	39.07	8	72.91

Table 2: Clast lithology (mean percentages) of tills at Weybourne Town Pit.

Unit	*Till*	*n*	*Flt*	*Clk*	*Sst*	*Cg*	*Lst*	*Irns*	*Mlrs*	*Mst*	*Shl*	*Qtz*	*Qtt*	*Ig/Met*
6	u. pale	1284	15.3	81.1	-	0.1	0.7	0.1	-	0.4	0.8	1.4	0.2	-
5	u. brown	99	61.7	17.3	0.9	8.2	2.0		0.8	-	1.9	6.1	1.5	0.5
4b	Dark pale	507	30.8	59.6	-	1.6	0.6	-	-	0.4	2.6	3.6	1.0	-
4ac	l. pale	1053	15.4	80.0	0.2	0.2	0.8	0.2	0.4	-	0.7	2.2	0.5	-
3	l. brown	388	72.4	2.8	0.5	2.8	1.0	-	1.6	-	7.0	8.8	2.6	0.3

Flt, flint; Clk, chalk; Sst, sandstone; Cg, Weybourne Crag (iron cemented quartz sand); Lst, limestone (undiff.); Irns, ironstone; Mrls, Mid. Jurassic orange marlstone; Mst, grey mudstone; Shl, shell fragments; Qtz, quartz; Qtt, quartzite; Ig / Met, igneous or metamorphic

Unit 3 is a yellowish-brown, (10YR 5-6/8) silty-sand diamicton up to 0.6 m thick with a strongly undulating outcrop in the section. The texture is similar to, but slightly coarser than, unit 2 with some small clasts and more sand. Counts of opaque and non-opaque heavy minerals from the 63μm fraction (Parker, 1997) also indicate an affinity between these two units. Between 3 and 4 m (Fig. 2) a thin layer of unit 3 sediment extends upwards into the overlying unit and tapers towards the right of the section, apparently eastwards. A very thin, discontinuous lens of unit 3-type sediment, extending from 0.8 m intersects this tapering feature which is interpreted as a drag fold. Constituent clasts are predominantly flint (72.4%), with subsidiary quartz, and shells. Very small quantities of chalk, iron-cemented 'Weybourne Crag' and quartzite and traces of marlstone, limestone, sandstone and igneous / metamorphic rocks are also present. The calcium carbonate content is relatively low (13.6%). Clast macrofabric analysis shows a bimodally clustered fabric trending NNW-SSE (Fig. 4A).

The contact of unit 3 with unit 2 is sharp and erosional and is interpreted as the plane of décollement at the base of a glacial sedimentary stack (cf. Banham, 1977; Hart and Boulton, 1991b). The obvious textural and mineralogical similarities between units 2 and 3 suggest that unit 3 is derived, partially, from unit 2, but its dominant flint clast component together with the cemented 'Crag' sand and low chalk content suggests that this unit is derived largely from local early Quaternary sand deposits overlying Chalk.

Unit 4 is a pale yellow (2.5Y7/4), silt-rich diamicton. From 0-4.3 m (Fig. 2) it is about 0.3 m thick but thickens abruptly to over 1 m thick across most of the rest of the exposure, except where a lens of very pale brown (10YR 7/3), slightly sandier diamicton occurs within it. This lens is not laterally traceable, appearing sharply at about 4.8 m, then gradually thinning to zero at about 6.8 m. The clast content of unit 4 is predominantly chalk (79.9%; $CaCO_3$ content about 65%) and flint (15.4%) suggesting a derivation largely from Chalk bedrock. There are also small quantities of quartz and traces of Mesozoic rocks including limestone, sandstone and carstone. The clast content of the pale brown lens (unit 4b) is intermediate between that of unit 3 and unit 4 with more flint (30.8%) and less chalk (59.6%) suggesting a degree of mixing of units 3 and 4. Both the upper and lower boundaries of unit 4 are sharp, but wavy or irregular, and it contains lenses and a thin discontinuous stringer (0.8-4.1 m, Fig. 2) of the adjacent sediment, again suggesting some intermixing.

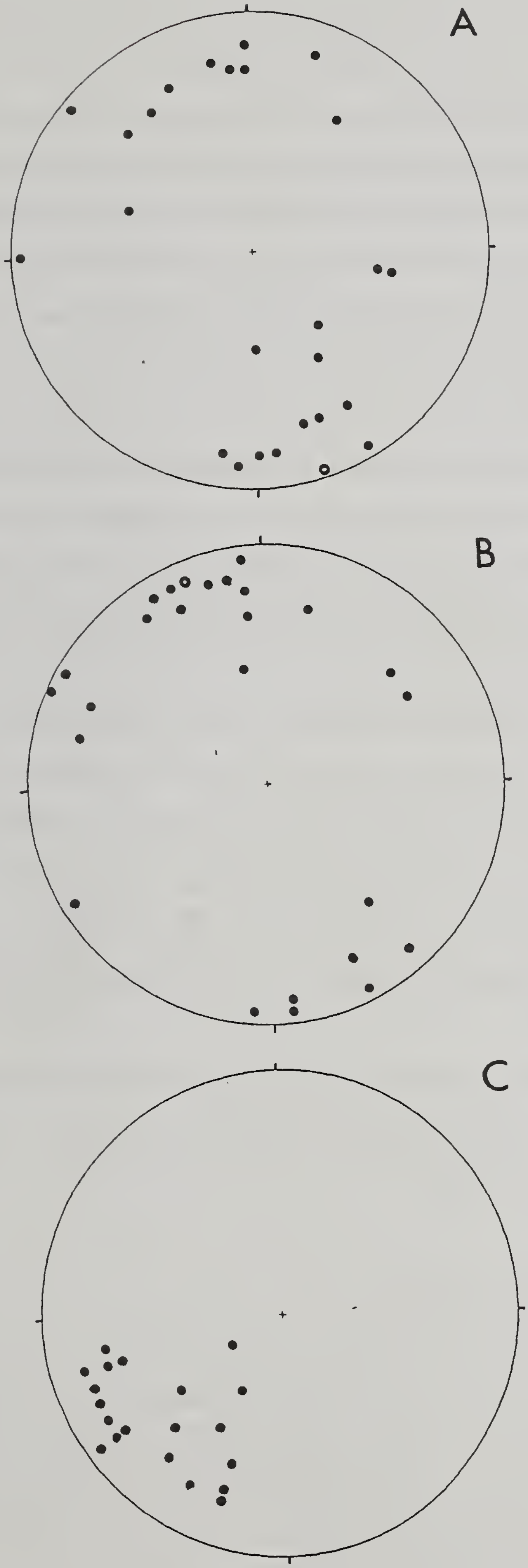


Fig. 4. Clast macrofabric analysis of units 3 (A) (n = 25) and 6 (B) (n = 25) and structural analysis (pole to plane) of tectonic folds (n=22). Circles in A and B represent resultant vectors.

Unit 5 is a thin (up to 40 cm, but more commonly 10 cm thick), brown (10YR 5/4-4/3), silty, sandy diamicton, except for a more gravely lens around the 4 m mark (Fig. 2) which extends downwards into unit 4. The clast content of unit 5 is dominantly flint (61.7%) with subsidiary amounts of Chalk (17.3%), cemented 'Crag' sand (8.2%) and quartz (6.1%). Traces of sandstone, limestone, marlstone, shell, quartzite and igneous and metamorphic rocks are also present. Unit 5 contains a thin lens of unit 3-type material, especially in the western part of the section. Beyond 8.2 m the unit bifurcates and the intervening space is occupied by a lens (unit 5b) of highly calcareous sediment with the appearance of deformed and reconstituted Chalk. Unit 5 has an undulating outcrop in the section reflecting that of unit 3.

In general appearance, colour, $CaCO_3$ content and clast composition unit 5 closely resembles unit 3. It is assumed that textural differences reflect a slightly different provenance. The high Chalk clast content and the highly calcareous lens (unit 5b) suggests closer contact with Chalk bedrock during its transport to this site than that experienced by unit 3.

Unit 6 so closely reflects unit 4 in almost all respects - colour, texture, $CaCO_3$ content, clast lithology and structure (it contains thin stringers of unit 3 and unit 5-type material), that it is considered to have the same origin as unit 4. Clast macrofabric analysis shows a clustered fabric trending NNW-SSE (Fig. 4B). An additional feature of unit 6, related to its high position in the sedimentary sequence, is the secondary nodules of $CaCO_3$ which are considered to be pedogenic in origin, being redeposited, dissolved Chalk from percolating ground water.

The orientation of the folds in unit 4 between 3 and 4 m and the dip direction of units 3 and 5 were measured. The results of the structural analysis clearly indicate stress from the south-westerly quadrant (Fig. 4C).

DISCUSSION

The sorting characteristics, erratic content and structural relationships of the sediments exposed in Weybourne Town Pit all indicate that units 3-6 represent tills, and that the boundary at the base of unit 3 constitutes the plane of décollement at the base of a deformed glacial sedimentary stack (Banham, 1977; Hart et al., 1990), emplaced subglacially by a land-based glacier (Hart and Roberts, 1994).

Analysis of colour, particle size, clast lithology and calcium carbonate content clearly indicates that two till facies are present at Weybourne Town Pit, represented by units 3 and 5, and 4 and 6, respectively.

The dark brown till (units 3 and 5) is dominated by locally derived Upper Cretaceous flint and Chalk, and what has previously been termed 'Weybourne Crag' (e.g. Reid, 1882), a ferruginous sandy, gravely sediment associated with the Cromer Forest Bed Formation, but the inclusion of Jurassic limestone and sandstone suggests a westerly provenance. High levels of quartz and quartzite may be derived from Lower to Middle Pleistocene deposits within the area of the North Sea (Cameron et al., 1992) or reworked from terrestrial, quartz-rich fluvial deposits which are widespread in East Anglia (Hey, 1976; Green and McGregor, 1978; Rose, 1989, 1992; Whiteman and Rose, 1992; Hamblin and Moorlock, 1995; Rose et al., 1999) but, equally, they may also be reworked from local outcrops of 'Weybourne Crag' (Banham and Ranson, 1965). The consistent, but minor, inclusion of igneous and metamorphic rocks suggests a more distant source, either in northern Britain or Scandinavia, but the number of clasts involved is so small that reworking from the fluvial or 'crag' deposits mentioned above cannot be ruled out. Although no direct evidence of tills of Scandinavian provenance has been found at the two sites in this study, current models explaining the deposition of the North Sea Drift Formation (Eyles et al., 1989; Hart and Boulton, 1991a; Lunkka, 1994) do suggest that Scandinavian ice reached north-east Norfolk.

The clast component of the pale yellow, silty till (units 4 and 6) is dominated by chalk (76.5 - 79.9%) and the matrix is highly calcareous (62.70 - 67.25% $CaCO_3$). The remaining clast content consists of about 2% quartz and quartzite, with traces of Jurassic material and 'Weybourne Crag'. No igneous or metamorphic pebbles were found. On the basis of this clast lithological data the provenance of most of this till appears to be very local, with only a small input of material from further west or north-west. The high content of chalk clasts and $CaCO_3$ in the pale yellow till, coupled with the paucity of other lithologies, are indicative of the Weybourne Town Member (Lewis, 1999). The till is similar to the Marly Drift, a very chalk-rich facies of the Lowestoft Till Formation, considered to be of local provenance, which extends across the northern part of East Anglia (Harmer, 1909; Solomon, 1932; Straw, 1965, Perrin, et al., 1979; Ehlers et al., 1987; 1991). The interbedding of these two tills may therefore have arisen in the following way. The brown till is a 'crag'-rich Scandinavian till which has been (a) over-ridden by an advancing British ice sheet, (b) partially eroded during excavational

deforming bed conditions (Hart, 1995), (c) glaciotectonically mixed with the British Weybourne Town Member till and (d) subsequently deposited during aggradational deforming bed conditions. Structural evidence from clast macrofabrics suggests deposition from the north-west. In contrast, fold measurements strongly indicate stress applied from the south-west. This may indicate that the clast macrofabric trend is transverse to the direction of ice movement, a pattern often associated with compressive stress (Boulton, 1987). Alternatively, the fold structures may represent post-glacial slumping (Banham and Ranson, 1965). A possible solution to these uncertainties concerning local provenance is provided by the analysis of the micropalaeontology of Chalk clasts in the tills.

CHALK MICROPALAEONTOLOGY

Chalk is an extensive bedrock in East Anglia and offshore, and has been subdivided into several biozones on the basis of its microfossils. It therefore provides an additional means of provenancing tills and reconstructing ice flow directions at a more local scale than that achievable using far-travelled erratics. The outcrops of Upper Cretaceous chalk zones in north Norfolk (Fig. 1), adopted for this study, are those mapped by Peake and Hancock (1961, revised 1978) and modified by R.N. Mortimore and C.J. Wood (personal communication). Weybourne Town Pit and the Weybourne cliff sections are currently located above basal Upper Campanian Chalk (Fig. 1).

The geomorphology of the preglacial landsurface is not known, but it is likely to have been higher than today, with an unknown thickness of chalk (possibly in excess of 20 m - see below) having been glacially eroded (Clayton, 1996; in press). As the dip of strata is about 1^{o} in the study area it is likely that Middle Pleistocene outcrop boundaries were an unknown distance west of present boundaries.

Currently offshore subcrops of chalk exist below many metres of Pleistocene sediments, where outcrops can be found occasionally at the base of basins (Balson and Cameron, 1985; Cameron et al., 1992). In pre-Anglian times much of this chalk was exposed, although adjacent Jurassic rocks were blanketed by early Pleistocene sediments (Cameron et al., 1992, Figs. 93 & 94).

Chalk clasts for micropalaeontological analysis were obtained from several layers (units 3-6) of both of the tills at the Weybourne Town Pit. Additional data were obtained from similar interbedded till units exposed in sea cliffs north of Weybourne Town Pit, where

they rest directly on Chalk bedrock (Banham, 1965; Banham and Ransom, 1965; Ehlers et al., 1987; 1991)

RESULTS

The presence or absence of specific species of foraminifera in chalk clast samples and calcareous matrix samples from the tills at Weybourne Town Pit and the Weybourne cliffs is presented in Table 3. The age range of each sample is summarised in Figure 5.

Clasts from the pale coloured till (units 4 and 6) at Weybourne Town Pit are of upper Santonian to Lower Campanian age with a provenance located towards the west (SSW-NNW) of Weybourne. Clasts from the upper brown unit (5) are of Lower Campanian age, also with a westerly provenance but less distant than that of the pale till. Clasts from unit 3, the lowest brown till unit, are of Upper Campanian age with a provenance that lies to the east of the quarry.

Till matrix data, in general, indicates a broader range of age, presumably reflecting comminution and mixing of different chalks during transport and deposition. With only two exceptions, sample ages range from middle Santonian to Upper Campanian provenance, indicating provenance dominantly in the west, but also to the east of the site. The two exceptions are the lower brown till sample (unit 3) which is again restricted to Upper Campanian, and the chalk-rich lens, unit 5b, with an age extending from Upper Campanian back to middle Coniacian.

Clast samples from the Weybourne cliff sections range in age from middle Santonian and Lower Campanian for the brown till to lower Middle Campanian for the pale till. They are therefore similar to those from the Town Pit in having a western provenance.

It is noticeable that none of the samples have a provenance which includes the Maastrichtian Chalk biozone. This is not surprising as this biozone is, apparently, entirely buried by more recent strata. However, chalk of this age does form part of large rafts in the Cromer Member of the North Sea Drift Formation near Sidestrand so was not entirely beyond the reach of glacial erosion.

Sample	Location	*Archaeoglobogerina cretosa*	*A. anglica*	*Bolivina incrassata crassata*	*Bolivinoides laevigatus laevigatus*	*B. decoratus decoratus*	*B. strilligatus*	*Cibicidoides voltziana*	*Gavelinella stilligera*	*G. monterelinensis*	*G. clementiana*	*G. tourainensis*	*G. cristata*	*G. usakensis*	*G. pertusa*	*G. eriksdalensis*	*G. lorieana*	*Globogerelliniodes aspera*	*G. rowei*	*Globorotilites micheliniana*	*Globotruncana bulliodes*	*G. plummerae*
CLASTS																						
1	E. cliffs brown till	X							X						X					X	X	
2	E. cliffs pale till									X	X				X					X		
3	W. cliffs pale till								X						X					X		
4	W. cliffs brown till								X													
s6	Town Pit Unit 6												X									
s4	Town Pit Unit 6												X									
t3	Town Pit Unit 6																					
5	Town Pit Unit 5b				X															X		
t2	Town Pit Unit 5a																					
6	Town Pit Unit 4								X									X			X	
MATRIX																						
4	Town Pit Unit 6	X		X		X			X	X			X	X	X			X		X	X	
11	Town Pit Unit 5b	X						X										X				
5	Town Pit Unit 5b								X	X			X			X	X		X	X		X
3	Town Pit Unit 5a									X			X		X			X		X		
8	Town Pit Unit 4b		X	X		X			X		X		X		X						X	
9	Town Pit Unit 4b			X			X							X						X		
7	Town Pit Unit 4a								X	X			X		X					X		
2	Town Pit Unit 4a								X		X				X							
1	Town Pit Unit 3										X									X		
20	Town Pit Unit 2	X		X	X	X			X			X			X					X		
22	Town Pit Unit 1c													X						X		
21	Town Pit Unit 1b													X								

Sample	Location	*yroidinoides nitidus*	*Hedberghella holmsdalensis*	*Loxostomum eleyi*	*Marsonella ozawi*	*Osangularia cordieriana*	*O. whitei*	*Plectina cenomanica*	*Praebulimina laevis*	*P. carseyae*	*P. obtusa*	*P. reussi*	*Pullenia quaternaria*	*Reusella szajanochae praecursor*	*Reusella szajanochae szajanochae*	*Rugoglobogerina rugosa*	*Stensioena exculpta gracilis*	*S. exculpta exculpta*	*S. granulata incondita*	*S. granulata perfecta*	*S. pommerana*	*Whitinella baltica*
CLASTS																						
1	E. cliffs brown till					X						X					X					
2	E. cliffs pale till	X				X			X								X				X	
3	W. cliffs pale till		X						X	X			X				X				X	
4	W. cliffs brown till											X					X		X			X
s6	Town Pit Unit 6					X	X			X												
s4	Town Pit Unit 6						X					X									X	
t3	Town Pit Unit 6																					
5	Town Pit Unit 5b															X					X	
t2	Town Pit Unit 5a																					
6	Town Pit Unit 4	X										X					X		X			X
MATRIX																						
4	Town Pit Unit 6	X							X					X			X		X			
11	Town Pit Unit 5b	X				X			X								X		X			
5	Town Pit Unit 5b			X						X			X				X	X				
3	Town Pit Unit 5a	X															X			X		
8	Town Pit Unit 4b	X				X				X							X				X	
9	Town Pit Unit 4b	X							X	X	X				X		X		X		X	
7	Town Pit Unit 4a				X					X			X				X		X			
2	Town Pit Unit 4a	X															X			X		
1	Town Pit Unit 3					X				X											X	
20	Town Pit Unit 2											X				X	X		X		X	
22	Town Pit Unit 1c	X								X		X	X									X
21	Town Pit Unit 1b																					

Table 3: Presence/absence of foraminifera identified in clast and matrix samples from Weybourne Town Pit.

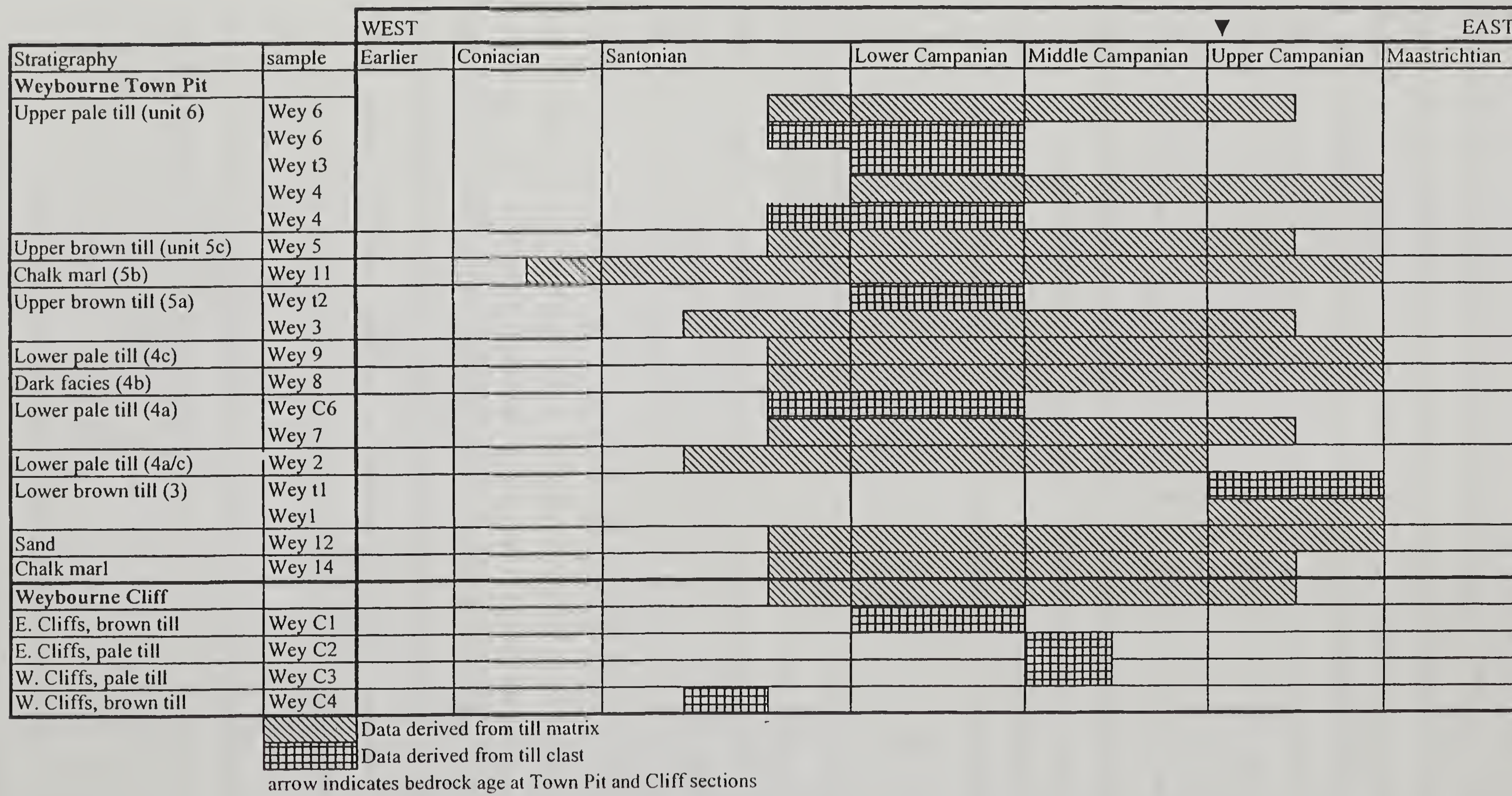


Fig. 5. Provenance of microfossils from Weybourne Town Pit (upper section) and Weybourne Cliffs (lower section). Note arrow at top of Figure represents position of Weybourne in relation to Chalk biozones.

INTERPRETATION OF PROVENANCE INDICATED BY CHALK MICROPALAEONTOLOGICAL ANALYSIS

Chalk clasts from tills exposed in the Weybourne sea cliffs clearly indicate a westerly provenance between south-west and north-west. This is consistent with south-west to north-east regional ice sheet flow directions, determined by West and Donner (1956) and Ehlers et al. (1987, 1991), which were responsible for the deposition of the lower units of Lowestoft Till.

The results from Weybourne Town Pit are less clear, especially those from matrix samples. Most of the chalk clasts have a westerly provenance, as do most of the matrix samples. However, both clast and matrix samples from the lower brown till (unit 3) and the majority of the other matrix samples possess a component with an easterly provenance. Two possible explanations of this situation are offered. In the first, two different glacial advances are envisaged, one from an easterly direction and the other from a westerly direction. In this case ice would initially have entered the area from an easterly direction, bringing highest Campanian chalks and depositing material including the lower brown till (unit 3) at Weybourne Town Pit. Subsequently, ice would have moved from a westerly direction, bringing older Campanian and Santonian rocks into the area, overriding, deforming and mixing material from the two advances together. Alternatively, it is possible to envisage a single advance, from a westerly direction, being responsible for the deposition of both tills at Weybourne and those of the surrounding area. In this model, the preglacial land surface, prior to glacial erosion, is assumed to be higher than today. With a regional dip of chalk strata of 1^{o} or less chalk zone boundaries would have been further to the west than their present position and the preglacial chalk outcropping at Weybourne Town could have been topmost Upper Campanian. Thus, an ice sheet approaching from a westerly direction could have eroded highest Campanian Chalk down to basal Upper Campanian chalk before depositing comminuted highest Campanian Chalk as the matrix of the tills, and giving the appearance of an easterly provenance (Fig. 6). Subsequent erosion of basal Upper Campanian and Upper Santonian strata, also from a westerly direction, contributed the larger clast component of the tills.

The thickness of chalk needed to be stripped off by glacial erosion to satisfy this scenario can be calculated using simple trigonometry (Fig. 7). The distance from the location of Weybourne Town Pit to the current upper boundary of Upper Campanian chalk is about 11 km (Peake and Hancock, 1978). Assuming Chalk strata dipping at 0.1^{o}, then the thickness of erosion required to bring lowest Campanian Chalk to the surface at

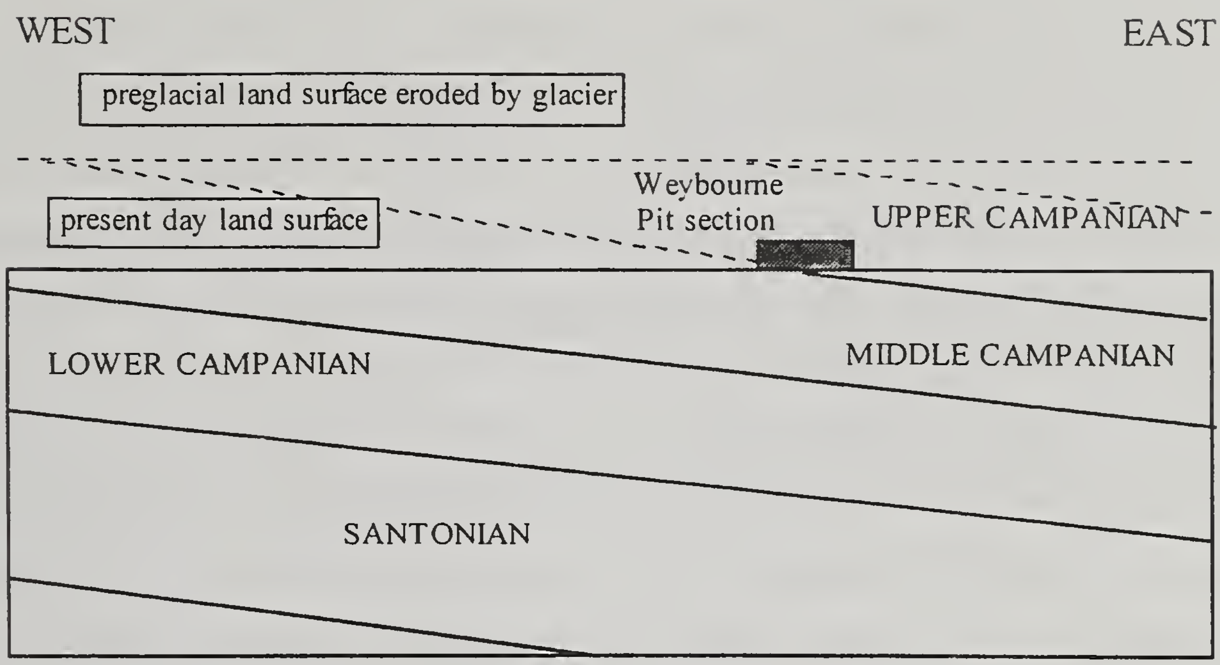


Fig. 6. Possible explanation of 'erroneous' chalk provenance data at Weybourne

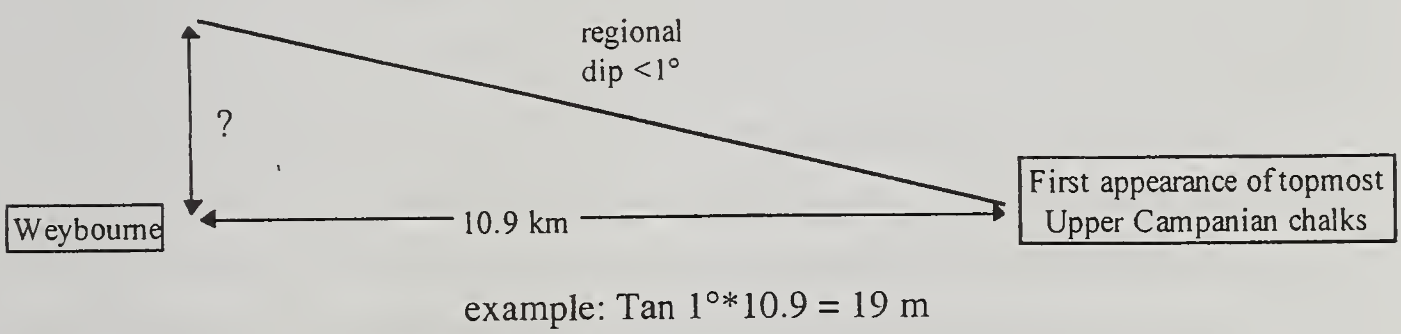


Fig. 7. Calculation of bedrock depth removal needed for single glaciation to form Weybourne sediments.

Weybourne Town would have been only 1.9 m. With a dip of 0.5°, the necessary erosion would have been 9.5 m and with a dip of 1°, 19 m (Fig. 7). This amount of glacial erosion is not difficult to envisage for relatively soft bedrock like Chalk. Clayton (in press) has calculated that the Chalk escarpment in the western part of East Anglia was probably lowered by as much as 49 m.

Some support for one or other of these reconstructions is available from the correlation of Weybourne tills with those of the North Sea Drift Formation (NSDF) in adjacent areas. The pale, yellowish till has already been correlated with the 'marly drift' of north Norfolk to the west (Ehlers et al., 1987, 1991). The brown till is unlikely to correlate with the Walcott Diamicton as it contains too little silt and $CaCO_3$, nor with the

Mundesley Diamicton (Mundesley-Trimingham Member, Lewis, 1999) as it shows no evidence of glaciomarine deposition (Lunkka, 1994). The closest comparison, in term of texture, structure and $CaCO_3$ content, between the brown till at Weybourne and members of the NSDF to the east and south-east is with the Happisburgh and Cromer Diamictons, Lunkka's (1994) first and third diamictons, which he attributed to deposition by Scandinavian ice advancing from the north-east and later, from the north-west. If these correlations are correct, Chalk of Upper Campanian and/or Maastrichtian age might be expected to be present in these NSDF tills and the brown till at Weybourne. However, although the Cromer Till contains large Chalk erratics, at least partially of Maastrichtian age, Maastrictian Chalk is absent from all samples analysed during this study. As the Happisburgh Member also contains no Maastrichtian Chalk it is the most likely correlative of the brown till at Weybourne.

SUMMARY OF PROVENANCE AND DIRECTIONAL DATA

Provenance and ice movement directional data relating to Weybourne Town Pit and adjacent areas is summarised in Table 4.

The brown till:

- is the lowest till in the sequence.
- contains lithological evidence of W-NW-N provenance.
- contains significant amounts of local 'Weybourne Crag'.
- contains a very small igneous/metamorphic fraction which has not been provenanced.
- contains equivocal micropalaeontological evidence of both eastern and western provenance.
- does not contain Maastrichtian Chalk, but that may not have been exposed.
- within unit 5 contains a marl lens with the oldest chalk analysed.
- matrix shows variable strength clast fabrics trending NNW-SSE.

These provenance and directional indicators contained in the brown till at Weybourne do not give a clear indication of the direction of flow of the ice sheet which deposited it. Although there is evidence of westerly or north-westerly provenance it also contains material which could have come from the north and east of Weybourne. In this study the brown till is correlated with the Happisburgh Member, the first evidence of the glacierization of north Norfolk. The Happisburgh Member is attributed by Lunkka (1994)

Table 4: Summary of provenance and directional indicators from sediments at Weybourne Town Pit.

Proxy	*Sedimentary units*	
	pale yellow till	*brown till*
Fabric ϕ * ° @	N -S / SW -NE / NNW-SSE	SSE - NNW / NNW-SSE
Structure ° ♣ @	SW to NE	SW to NE
Clast microfossils @	NNW to SSW	NNW to SSW and NNE to SSE
Matrix microfossils @	NNW to SSW and NNE to SSE	NNW to SSW and NNE to SSE
Macrofossils °	NNE to SSE	
Clast lithology ° @	NE to E	NW to W (and E?)

Key, ϕ after West and Donner, 1956; * Ehlers et al, 1987; ° Banham and Ranson, 1965; ♣ after Banham, 1965; @ this study

and others (Reid, 1882; Banham, 1971) to Scandinavian ice flowing from the north-east, on the basis of clast fabrics and structure, and by Hart and Boulton (1991a) to Scandinavian ice flowing from the north or north-west on the basis of structural data. Therefore, the brown till could be assigned to the North Sea Drift Formation, on the basis of its chalk microfossils and possibly its igneous/metamorphic clast lithologies. However, other evidence does not support this NE provenance, as the till also contains chalk microfossils and clast lithologies of a westerly derivation and structural and fabric evidence pointing to a south-westerly derivation. This equivocal evidence, coupled with its stratigraphical position and close structural relationship with the pale, yellowish till suggests that the brown till has experienced post-depositional modification in the form of deformation and mixing.

The pale yellow till:

- is interbedded with the brown till and contains thin lenses and drag folds composed of that till.
- contains lithological evidence indicating provenance from a westerly or north-westerly direction; in particular contains significantly more chalk than the brown till.
- lacks an igneous/metamorphic component suggesting that the ice which deposited it has not travelled from a distant source in northern Britain or Scandinavia.

- contains micropalaeontological evidence of both eastern and western provenance, though the west direction seems to dominate. There are no clasts of an easterly provenance, only matrix.
- contains clast fabric evidence indicating ice flow trending NNW-SSE, NE-SW and N-S.
- contains structural evidence for ice movement from the south-west.
- is correlated with the Marly Drift which occurs generally to the west of Weybourne.

The pale, yellowish till possesses stronger indications of a westerly provenance than the brown till, although there is also Chalk from the east in its matrix. However, the structural and clast fabric characteristics, and its general similarity to the 'marly drift' is strong evidence for ice movement from a south-westerly to westerly direction.

It is possible that both tills can be attributed to a single advance of a British ice sheet. Fabric, structure, chalk micro- and macrofossils and clast lithologies all agree with a south-western provenance. However, this assumes that the igneous/metamorphic clasts are of British origin, and it is difficult to explain why they are only found in the brown till. A second difficulty is the need to remove up to 5 m of 'Weybourne Crag' (Banham and Ransom, 1965) before eroding the Chalk bedrock and depositing 'crag'-rich, brown till on the eroded chalk surface. It is difficult to envisage this being achieved without fully mixing 'Crag' and Chalk together. Furthermore, this model assigns no evidence to glacierization by Scandinavian ice, which Hart and Boulton (1991a) and Lunkka (1994) believe affected the Weybourne area.

These difficulties can be removed if deposition of the tills involves more than one ice sheet and some mixing of till components occurs. In this model, a Scandinavian Ice Sheet initially reached the Weybourne area from a northerly to north-easterly direction and deposited a till composed primarily of 'Weybourne Crag' with a few igneous and metamorphic rocks and Chalk micro-fossils with an easterly provenance. A subsequent advance of a British ice sheet from the south-west remobilized the Scandinavian material under excavational deforming bed conditions (Hart, 1995). During this process it is possible that Chalk, which now has an easterly provenance, but then occurred to the west of Weybourne, was mixed with till from the British ice sheet, while some of the western lithologies were mixed into the brown till under conditions of high pore-water pressure and low material strength. Under conditions of compressive stress the brown till was

sheared and folded by the British ice sheet, resulting in interbedding and boudinage structures (Banham, 1988; Banham and Ranson, 1965) and the development of transverse clast orientation. The overall result is a palimpsest dispersal pattern (Parent et al. 1996).

CONCLUSIONS

The Weybourne area is clearly important to studies of glaciation in North Norfolk, and in East Anglia in general, as it is one of the few sites where evidence of both British and Scandinavian ice sheets exists. Sedimentary, structural and palaeontological evidence indicates that a British Ice Sheet moving from a south-westerly direction deformed existing till deposited by a Scandinavian ice sheet. The tills at Weybourne are considered to be equivalent to the Happisburgh Member of the North Sea Drift Formation and the 'marly drift' of the Lowestoft Formation. Finally, understanding the sequence of events at Weybourne has been greatly facilitated, for the first time, by the use of Chalk micropalaeontology for provenancing tills.

ACKNOWLEDGEMENTS

CW is grateful to the Geography Division at the University of Brighton for the studentship awarded to PRF. Funding from BGS/NERC university collaboration fund is also acknowledged. Thanks are due to Haydon Bailey for the analysis of samples from units 1 and 3 at Weybourne, carried out during preliminary tests of the micropalaeontological technique and to Helen Parker for supplying unpublished heavy mineral data. Rory Mortimore and Chris Wood kindly updated the Chalk microfossil zones of Peake and Hancock. BSPM, RJOH and IPW publish with permission of the Director of the BGS.

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[Manuscript received 27 October 1999; Revision accepted 17 February 2000]

EROSION BY ICE, RIVERS AND THE SEA: A COMPARISON OF RATES OF EROSION AND TOTAL DENUDATION ACHIEVED IN EAST ANGLIA OVER THE LAST HALF MILLION YEARS

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ABSTRACT

Erosion rates and total erosion achieved in East Anglia over the last 450,000 years (since the end of the Cromerian) are described. Processes include glaciation (the Anglian ice advance), rain and rivers ('normal erosion' of W.M. Davis), solution (part of the total work done by rain and rivers) and marine erosion. The data illustrate well the concepts of magnitude and frequency, whilst quantification corrects the errors of estimates based on theory (as represented in the literature) or the perception derived from observations of process and landforms today.

INTRODUCTION

Over the last few years, I have completed work on three aspects of geomorphological change in East Anglia in the later part of the Quaternary. The purpose of this paper is to summarise each of the three examples, which are published in full elsewhere, and to compare them in order to get an improved understanding of the relative contributions of each of these styles of denudation to landform change in East Anglia. The glacial example concerns the work done by the Anglian glaciation, generally acknowledged to date from about 440,000 yr BP, whilst the fluvial example involves the dissection of an extensive, low plateau of Anglian till, i.e. the work done by 'rain and rivers' since the end of the Anglian glaciation. Inevitably, the marine example is limited to the latter half of the Holocene, the last 5-6,000 years since sea level has stood at its present level. Suitable allowance can be made for earlier interglacial periods with similar high sea levels. In addition, earlier work by Edwards (1973) on the removal of material in solution is included in the comparative tables.

Bull. geol. Soc. Norfolk (for 2000) **50**, 47-69 (Published 2000)

The measurement of denudation over time will always involve considerations of magnitude and frequency. This is particularly true when different processes are being compared. We still remain unsure whether there was a single ice advance to produce the Anglian glacial deposits of East Anglia, but, whether there were two advances or one, they will have occupied little more than ten or twenty thousand years of the last 440,000. Magnitude values are high, frequency is very low as the area lies beyond the limit of ice advance of glaciations younger than the Anglian glaciation. In the case of fluvial erosion (including solution), the rates of operation are relatively slow, especially in the gentle relief characteristic of East Anglia, but the process has continued unabated over the whole of post-Anglian time. Marine erosion has the additional complication that while it is an essentially continuous process, the location where it is effective changes greatly as sea level rises and falls. Over the last 450,000 years, it will have been close to its present level on four occasions, i.e. during each of the four interglacial periods since the Anglian ice advance. It is unlikely that any of these had a stable, high sea level for more than 10,000 years, so again, so far as our coastal zone is concerned, magnitudes are relatively high, frequency is low.

A further complication is that whilst the processes of solution, the entrainment of sediment by rivers and marine erosion generally lead to the removal of almost all the resulting sediment from the present land area, glaciation has produced large volumes of till which still mantle much of the area, although steadily being dissected by 'normal' erosion.

GLACIAL EROSION DURING THE ANGLIAN STAGE

It is widely, though perhaps not universally, accepted that the great embayment of eastern England represented by The Wash and The Fens is the work of an eroding ice sheet. Geomorphological arguments for this include the general form of the embayment, the size of the breach in the Chalk escarpment represented by The Wash, and the form of the Chalk escarpment. The escarpment is high in the Chilterns, with a crest close to the outcrop of the basal chalk at the foot of the scarp, and is low in Cambridgeshire, Suffolk and Norfolk, with the crest set back to the east. Geological argument rests in the main on the area of till (the Chalky Boulder Clay) which lies around the basin and which lithologically supports erosion of the Jurassic clays and the Chalk (Harmer, 1909; Perrin *et al.*, 1979) and which reflects in the orientation of the clasts within the till an outwardly-radiating movement from the Fens/Wash embayment (West & Donner, 1956). There is

other geological evidence; notably the evidence of Woodward, (1903, and see also Bromley, 1966; Hopson, 1995) that the chalk outcrop in the Royston area has been greatly disturbed with displacement of huge blocks of chalk revealed by careful geological mapping and by several sections showing displacement or folding of strata. Woodward (1903, p.369) concluded "No other explanation is reasonable, and in my opinion none other is possible, than that the Chalk was disturbed by the agent which formed and distributed the Chalky Boulder-Clay."

The link between the pattern of ice movement, and the lithology of both the eroded areas and of the resulting till surrounding the basin was confidently set out ninety years ago by Harmer. He wrote of the 25-mile wide gap in the Chalk escarpment between Swaffham and Newmarket "Through this gap, which was no doubt deepened as well as widened in the process, the ice seems to have poured in great volume from the Fenland, fanning out thence to the east and south-east, covering an area of something like 1,000 square miles with Chalky-Kimeridgian boulder-clay which approaches in places 100 ft. in thickness, its surface reaching a maximum elevation of more than 300 ft. above sea-level. Only one possible source can be suggested for this enormous mass of Kimeridgian detritus. The Fenland to the west, and near at hand, was the quarry from which it was dug; the accumulation of material in the one area being the measure of its removal from the other" (Harmer, 1909, p.114). Today, we would add that the presence of a deformable bed in the Jurassic mudrocks aided the development of this very active ice lobe and its ability to excavate the Fens/Wash basin.

The modification of the form of the Chalk cuesta was described in detail by Linton (1963) in his Presidential Address to the Institute of British Geographers as follows. "In the Chiltern Hills, which are believed never to have been overrun by the ice, the escarpment is a bold feature of very distinctive form, with characteristically a lower bench fronted by the escarpment of the Melbourne Rock, and set back from the latter, an upper scarp capped by the Chalk Rock that runs forward in a series of bold salients flanked by wing-like spurs." To the north, "the main escarpment of the Upper Chalk is replaced by a gentle, partly drift-covered rise to the low boulder-clay plateau of Suffolk. This is set back by many miles from the line of the escarpment farther south-west, and may be traced as a 'height of land' toward Bury St. Edmunds, sometimes nearer the south-eastern than the north-western limit of the Chalk outcrop. For upwards of thirty miles in north-west Suffolk and south-west Norfolk even this fails and the Fen streams Lark, Little Ouse and

Wissey head in an imperceptible divide some twenty-five miles east of the Gault outcrop. Around Swaffham the scarp feature reappears and may be followed, paralleled at a small distance by the Peddar's Way, to the sea near Hunstanton."

Perrin *et al* (1979) have described the lithological uniformity of most of the chalky boulder clay and shown the way in which a lobe of ice spilled up and out of the Fens to deposit the chalky boulder clay on the surrounding cuestas. Their Figure 10 (Fig. 1) shows an outwardly radiating pattern of ice movement, and their description of the lithology of the tills matches their local origin from erosion of the Chalk and the underlying Jurassic clays. The pattern of movement had of course been adduced earlier by general observation and was mapped with some precision by West & Donner (1956) using the orientation of clasts included in the various tills.

Quantifying the amount of erosion of the Chalk escarpment

The approach to estimating the volume removed by modification of the form and position of the Chalk escarpment, was to compare the unglaciated form (using the mean elevation of each National Grid kilometre square - Clayton & Shamoon, 1998a) of the Chilterns with the subdued feature to the north. The comparison was done by averaging the values for sample zones each 14 km wide along the strike in order to allow for the pattern of valleys and intervening ridges found everywhere, but particularly marked in the deep dry valleys of the Chilterns. The values were summed sector by sector (Fig. 2), since modification increased from near Royston to the Norfolk border and then declined again to the north, except for the total breach now occupied by The Wash. The details are described elsewhere (Clayton in press) and the shift in position of both the base of the escarpment and the water parting on the Chalk is shown in Figure 2. The values obtained from comparison of computer-stored values of mean altitude for kilometre grid squares were for mean volume removed per kilometre of scarp length, divided into Chalk and the underlying mudrock, mainly of Jurassic age. From the values computed from the length along the strike of each zone, a total volume for Chalk is obtained as shown in Table 1. The volume of mudrock derived from erosion of the Chalk escarpment is also indicated in Table 1.

Given all the approximations involved in this reconstruction a value for the volume of glacially-eroded Chalk of 243 km^3 will serve as the basis for further discussion and

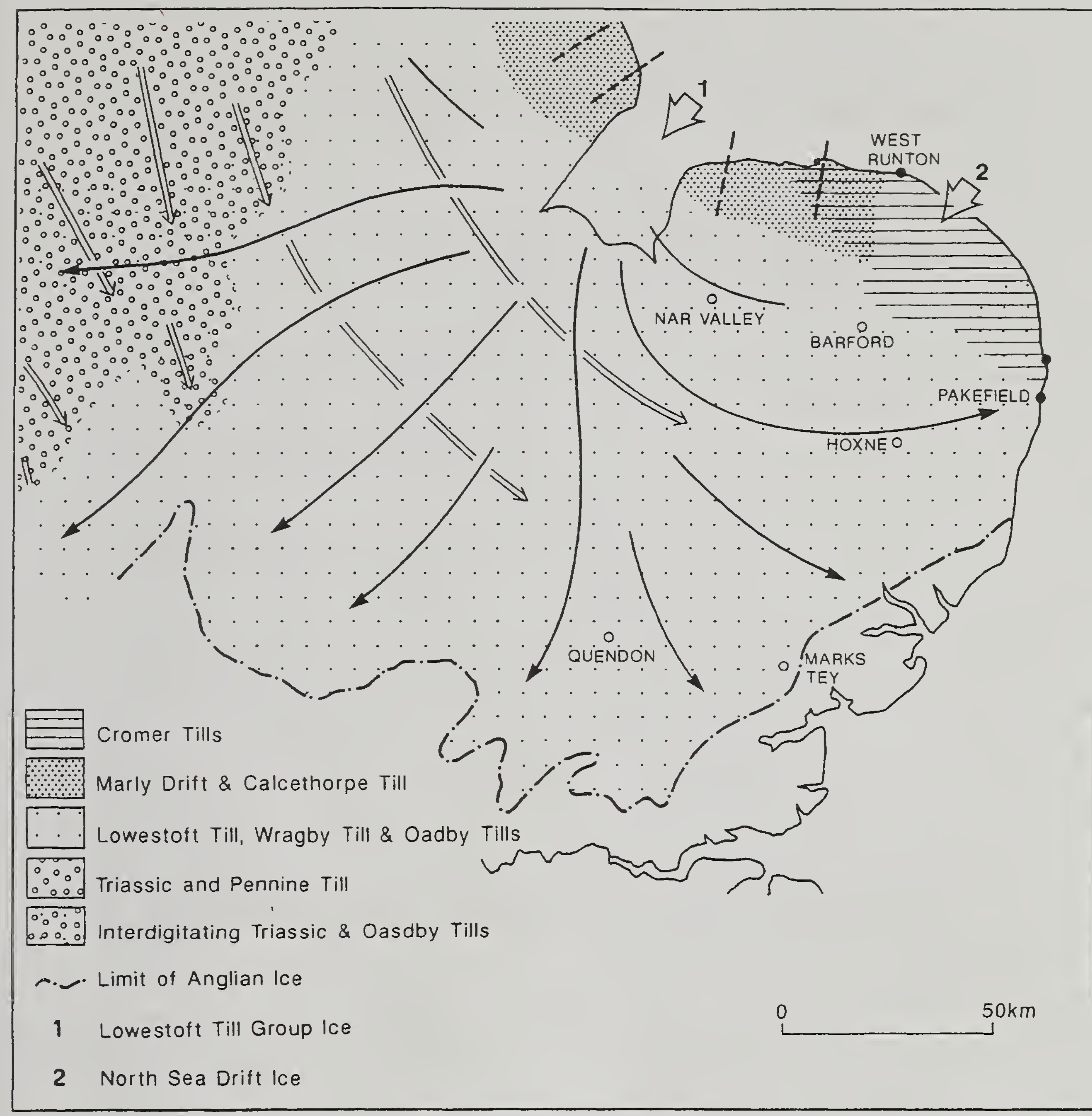

Fig. 1. Glaciation of the Fens/Wash basin and emplacement of the Chalky boulder clay of Anglian age.

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analysis. The 83 km³ of mudrock eroded from the escarpment zone forms only part of a larger volume principally excavated from the Jurassic outcrop from the Wash and Fens.

Glacial erosion of the Fen Basin

Calculation of the volume eroded from the weaker rocks outcropping west of the Chalk escarpment requires reconstruction of the pre-Anglian landscape of what is now The Fens. As Rose (1987, 1989; Bateman & Rose, 1994) has shown, the Fens were crossed from

Fig. 2. Glacial erosion and the reshaping of the Chalk escarpment in East Anglia.

west to east by the Bytham River which joined the Thames somewhere near Lowestoft. Immediately prior to the Anglian glaciation, this river flowed across what are now the Fens, entering near Castle Bytham (east of Melton Mowbray) at a height of 42 m and leaving near Bury St. Edmunds at about 16 m OD (Rose, 1987, Lewis & Bridgland, 1991). This gives a mean elevation for the valley floor across the Fens of 29 m.

Table 1: Calculation of eroded volumes of mudrock and Chalk; Anglian glaciation of Fens and Wash

	Area	Mean depth eroded	Volume
	(km^2)	(m)	(km^3)
Mudrock			
Fen Basin	3848	59.5	229.0
Chalk escarpment			
	3696	22.3	83.4
Total mudrock			312.4
Chalk			
Chalk escarpment			
	3696	65.7	242.9
Totals	7544	73.6	555.3

Table 2: Total original volume of sediment around the Fens/Wash embayment.

	Area	Mean thickness	Volume
	(km^2)	(m)	(km^3)
Area C	5464	20.0 (1)	109.3
Area D	7904	24.65 (2)	194.8
With allowance for solution (4m)			
Area C	5464	24.4 (3)	133.3
Area D	7904	29.05 (3)	229.6
Total Lowestoft till	13,368	26.7	362.9

(1) Thickness for Area C based on 60% at 25 m average thickness, plus thinning edge 40% at mean of 12.5 m.

(2) Thickness for Area D based on 70% at 29 m average thickness, plus thinning edge 30% at mean of 14.5 m.

(3) Current mean thickness adjusted by an allowance for solutional loss of 4.4 m depth over last 440,000 years.

We then need to estimate the overall mean level of the area of the Fens at that time. The Thames Basin above Goring Gap is developed on similar rocks, and on the whole without much change attributable to ice, though the northern part was glaciated and till mantles the higher interfluves. The Thames Basin is even today at a higher overall elevation than the destroyed Bytham Basin, since it enters the Jurassic vale at 78 m and leaves at Goring Gap at 44m, the average of 61m being 32 m higher than the average for the Bytham. Excluding some slopes rising to the Chalk margin, the average elevation of the Thames Basin between the Cotswold limestones and the Chalk escarpment (using digital data of values of mean altitude by kilometre grid squares for 1701 km^2) is 81.5m, i.e. 20.5 m above the mean elevation of the river.

Applying the difference between the mean elevation of the River Thames and the mean elevation of its drainage basin on the Jurassic mudrocks to the Bytham River gives an average elevation for the Fens at that time of 49.5 m. This figure was increased by 10 m to allow for the (poorly-known) elevation of the solid rocks below the Quaternary infill of the Fen Basin. Thus we use a value of average lowering of 59.5 m for Area A on Figure 3, an area of 3848 km^2 including the relevant part of The Wash; giving an eroded volume of mudrock of 229 km^3. Adding this volume to the mudrock eroded from the former Chalk escarpment, gives a total volume (Table 1) of 312 km^3 (Clayton, in press).

It is noteworthy that, despite using such different approaches to the calculation of eroded volumes for the two areas (Fen Basin and Chalk escarpment), the mean values for average eroded depth are of similar order; 59.5 m for the Fen Basin (Area A of Figure 3) and 88 m for the escarpment (B on Figure 3).

The volume of glacial sediments deposited around the Embayment

A large part of the sediment eroded from the area covered in Table 1 has been deposited around the basin. An estimate of the original volume (Table 2) gives a figure of 304 km^3. This employs current maximum thicknesses for the till of 29 m in Suffolk and Essex and 25 m west of the Fen Basin, together with an allowance for a 'feather edge' to the till sheet as explained in the notes to Table 2. As will be seen from the discussion of the work done by 'normal' erosion below, some further allowance is required for the lowering of the surface of the Chalky till as a result of solution over the past 440,000 years. A value of 4.4 m, the same as that used in calculating the long-term rate of normal erosion is therefore added to these mean till depths, resulting in an original volume of some 363 km^3.

Fig. 3. The areas utilised in the calculation of the quantitative impact of denudation on East Anglia over the past 450,000 years.

Since some glacial sediment will have been flushed out in solution or as outwash into the North Sea, this matches the estimate of eroded volume very well. In addition, although bulk data are sparse, there is also a close match between the proportions of Chalk (and flints) and mudrock for the eroded volumes and the Chalky till which surrounds the basin. For fines (<2 mm), the ratio (Perrin *et al.*, 1973) for Suffolk tills is 57:43

(mudrock:chalk), or 1.33 to 1, compared with the eroded ratio (Table 1) of 56:44, or 1.27 to 1. All this increases confidence in the results for the eroded volume. However, it must be noted that the effect of the deposition of glacial sediments on land is to reduce the net erosion achieved by the ice to some 192 km³ (Table 5).

Rates of glacial erosion

One implication of the erosional volumes is that the mean depth of erosion by the Anglian ice advance on mudrock and chalk together was of the order of 68 m (Table 1), a remarkable figure for what is currently thought to be a single ice advance lasting at most a few tens of thousands of years. By comparison, in the same post-Anglian period the Thames at Goring Gap incised about 55 m, though by no means all its catchment was lowered by the same amount. Even if future work establishes two separate glaciations, the mean depth of erosion by ice remains impressive and of the order of 3680 Bubnoffs (B, see below) and 460 B for the net erosion achieved over the area of Figure 3. Even when calculated as an average rate over the entire period since Isotope Stage 12 (440,000 yr BP), this is 164 B, and even the net erosion by ice represents a long-term rate of 20 B.

The term Bubnoff is a convenient one for representing rates of erosion, although it is still not widely used. Numerically, the value 10 B represents the removal of 10 mm from the entire surface over a period of 1000 years, or 10 m over 1 Ma (million years). However, the value may also be expressed in terms of volume removed from a unit area, such that 10 B also represents 10 m^3 km^{-2} yr^{-1}.

'NORMAL' EROSION OF THE CHALK DIPSLOPE IN SUFFOLK AND NORTHERN ESSEX

The area of the main till sheet (Area D of Figure 3) lying on the south-eastward-slope of the dip slope of the Chalk was outlined, omitting that part draining northwards towards The Wash. Generalised contours were drawn at 15 m vertical interval for the till surface by joining with a smoothly curving line the highest contours on each of the main ridges separating the rivers. These interfluves are broad and are believed to lie very close to the original depositional surface of the till. Thus while the generalised contours must necessarily be drawn boldly across each of the river valleys without any local evidence of their correct position, their position and orientation is strongly constrained by the

contours on the broad interfluves, and two or more operators following the same assumption that the interfluves were close to the original till surface would produce very similar maps. Further, the low slope of the reconstructed surface (at an average of 2 m/km) means that any disparities in the location of the reconstructed contours of the till surface have little effect on the final map of denudational lowering.

The depth of incision below this reconstructed surface may be determined at every point where the contours of the present land surface cross the reconstructed contours. Additional points may determined if required by interpolation between the contours (both of the present map and of the reconstructed surface). These points were marked across the whole area of interest and through them contours of the difference between the initial and the present surface were drawn. These have a consistent pattern from one river valley to another as will be seen from Figure 4. The modal depth of erosion measured over the whole area is 10.46 m. The greatest depths of incision reach almost 60 m (cf. the Thames at Goring Gap: 55 m), and this maximum value is approached in several of the main river valleys. Because the rivers slope more steeply than the initial surface in their upper reaches and more gently in their lower portions, the contours show greatest depth of downcutting midway down the slope of the surface.

The contribution of solution to total 'normal' erosion

Reflecting on the absence of surviving interglacial soils, we should also note that invariably the soil formed on the chalky boulder clay is free of chalk, so some lowering by solution has undoubtedly occurred. Solution is concentrated at the soil/rock interface and probably increases downslope as water moves under the influence of gravity in the surface regolith, so we may reasonably regard it has having operated over the entire surface. Contemporary values from Norfolk rivers (Edwards, 1973) are annual rates of removal of calcium carbonate (allowing for rainfall inputs) of 58.0 tonnes/km^2 for the Yare and 39.2 tonnes/km^2 for the Tud. These translate to rates (for $CaCO_3$ alone) of 22 and 15 B. Long-term values are probably lower, so a value of 10 B is taken, implying overall surface lowering by solution of 4.4 m.

In the circumstances this figure seems a reasonable basis by which to adjust the Suffolk data. The effect of this addition (to both the average depth of erosion and to the maximum depth of incision) is to increase the rate of denudation (see next section).

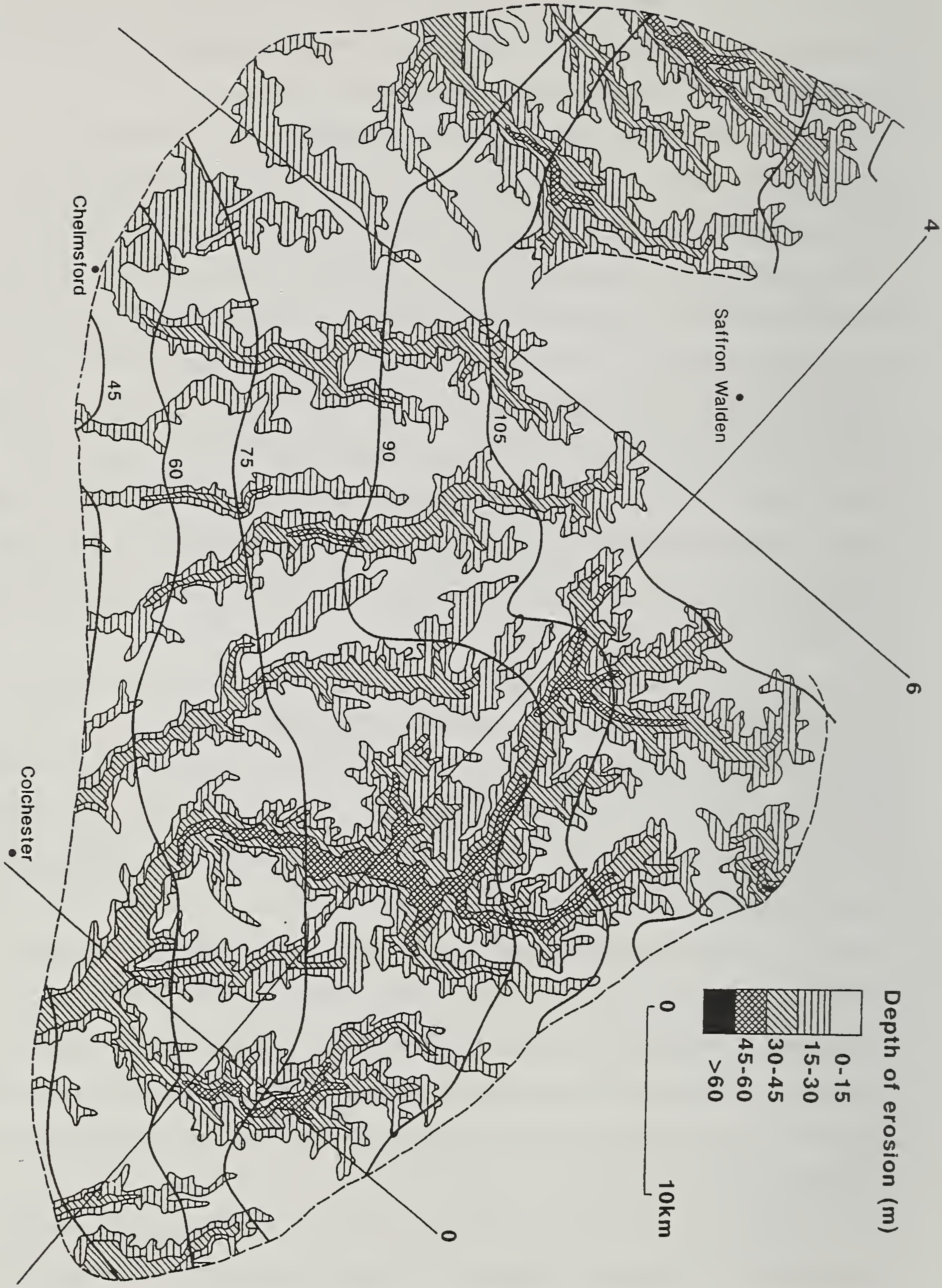

Fig. 4. The dissection of the Anglian till of Suffolk.

Table 3: Erosion of the Anglian Till in Suffolk

	Original measurements	With solutional lowering of 4.4 m
Number of data points	2325	2325
Average depth of erosion	16.91 m	21.31 m
Modal depth of erosion	10.46 m	14.86 m
Maximum depth of incision	59.6 m	64 m
Max. rate of incision (440ka)	135 B	145 B
Rate of erosion (440ka)	38.4 B	48.4 B

Total 'normal' erosion rates as measured by dissection of the Anglian Till

The evenly-spread effects of solution will have affected the whole of the chalky till, so that the reconstructed surface level is some 4.4 m below its original elevation. Thus 4.4 m should be added to the values (see Table 3) to allow for this. Then the maximum depth of valley incision (64 m) represents a time-averaged rate of incision of 148 B, using the value of 440 000 yr (Stage 12) for the Anglian glaciation. The areal average rate of denudation is much lower than the calculated time-average rate of valley incision, as the interfluves have been little reduced below the original surface. Sampling the pattern of Figure 2 on a regular (1 km) grid with 2235 points gives an average lowering of 16.91 m, or 21.3 m when solution is allowed for. Using the same Stage 12 date of 440,000 yr, this gives 48.4 B as the long-term rate of normal erosion (Table 3).

MARINE EROSION

Over the last 6000 years or so since the postglacial (Flandrian) rising sea level came close to the current level, the sea has been eroding the Norfolk cliffs. We have evidence of the contemporary rate of erosion based on map comparisons over the last hundred years, and the average is just under 1 m/yr (Cambers, 1976). It has been calculated that this results in the erosion of some 750,000 m^3 of sediment a year from a length of cliffs of about 40 km (Clayton, 1989). Contemporary erosion is well below this rate because of the considerable length of defended coast with various styles of built defences (Clayton, 1989). There seems every reason to believe that long-term erosion has been at rates close

to the average for the last 100 years, since wave energy has been maintained during persistent coastal retreat by offshore deepening. We know that erosion rates have been high for 1000 years or so, for several villages mentioned in Domesday have been lost to the sea. If we extend the rate back for 6000 years, we imply a shift in the position of the coast of some 6 km, and the offshore water depth at this distance offshore is now about 10 m. This implies offshore lowering at an average rate of about 1.5 to 2 mm/yr. Some small part of this has been contributed by recent sea-level rise, but the dominant long-term effect must be erosional lowering of the sea floor driven by wave energy and tidal currents, no doubt aided by bioerosion and perhaps some degree of submarine weathering of the Chalk.

Support for the view that this erosional system has persisted for such a long time is found in consideration of current and recent sediment balance. The erosional system persists because sediment is quickly removed from below the cliffs, mud and some 25% of the sand directly offshore; the remainder of the sand and gravel by longshore transport in the beach system southwards. Since 1935, Yarmouth North Denes has built out at 6 m/yr, testimony to the delivery of sediment from the north. Another confirmation of the efficiency of the sediment removal system is that the highest cliffs (c. 40 m) retreat the most rapidly; i.e. they fail to supply enough sediment to slow the system down. The calculation in Table 4 includes the Suffolk cliffs, which contribute a further 130,000 m^3/yr of sediment (Clayton et al., 1983). As will be seen from Table 5, the cliffs have contributed an estimated 5.4 km^3 in the past 6000 yr, the lowering of the sea floor a further 1.125 km^3, a total of 6.5 km^3.

Analogous to the deposition of till and glacial sands and gravels by the Anglian ice sheet, is the transport of eroded sediment from the active coast into mudflats, salt marshes and offshore sand banks. Only part of these are formally within the land area, and as the total volume is small, and deposits from previous interglacial stages have not, on the whole, survived, no adjustment is made to the figures presented here. However, it is necessary to make allowance for the fact that the present interglacial is not unique: sea level will have been high and close to the contemporary level in each of the four interglacial periods since the Anglian glaciation. Assuming that the average length of an interglacial is 10,000 years, this gives a total duration for marine erosion of the present land surface of 36,000 yr.

Table 4: Holocene marine erosion on the East Anglian cliffed coast

Cliff erosion				
mean rate (m/yr)	duration (yr)	mean height (m)	length (km)	volume (km^3)
0.9	6000	20	50	5.4
Sea-floor lowering				
	mean depth (m)	mean width offshore (km)	length (km)	volume (km^3)
	5.0	4.5	50	1.125
Total				**6.525**

The above total represents a maximum of 6000 years of Holocene time. There have been three post-Anglian interglacials before the present interglacial. If we assume that each has lasted 10,000 yr, then the total time for erosion at about the level of the present coast becomes 36,000 yr. Thus the total marine erosion since the Anglian glaciation becomes 39.1 km^3.

Extrapolation of the values from Table 4 to a total duration of 36,000 yr gives a total attributable to marine erosion of some 39.1 km^3, as shown in Table 5.

COMPARISON OF THE EROSIONAL WORK DONE

In comparing each of the erosional processes we have two main measures, the rate of erosion and the total volume removed. The relationship between the two involves estimation of the areas affected and the time for which the process has operated. A further complication is that glacial erosion deposited an estimated 70% of the eroded material on land, thus raising the surface level. The net effect of glacial erosion is therefore less than the gross volume of material removed: all the other processes of erosion remove the eroded material into the North Sea basin. The situation that glacial erosion is offset by the till volume left behind by that ice-sheet is unique to this part of the British Isles. Nowhere else are there extensive till deposits of this age, whilst further north the till volumes remaining on the landscape are offset by erosion over more than one glacial episode (Clayton, 1996).

Rates of erosion are summarised in column 1 of Table 5. The estimated area affected and the total time of operation since the Cromerian (here defined as 450,000 yr BP) are in columns 2 and 3, and the total volume eroded is then calculated in column 4, which is the measure of magnitude.

It is a remarkable coincidence that the rate of glacial erosion (calculated on the very broad assumption that it occupied 20 ka) is almost the same as the rate of marine erosion during each interglacial; thus the rate of erosion of the Norfolk coast matches in speed the rate at which the ice sheet was able to carve out the Fens and the Wash. The areas involved are, however, very different, an estimated 350 km² for coastal erosion, compared with the 7544 km² lowered by the advancing ice. Even allowing for deposition, the net glacial rate remains relatively high at 460 Bubnoffs, whilst the long-term rate of marine erosion is 248 B for the small area directly affected. In terms of rate of operation, normal erosion proceeds at only 1.3% of the gross rate of glacial erosion, and at 10.5% of the net rate when the effects of glacial deposition on land are allowed for.

The gross volume eroded by the processes considered here is some 990 km³; after allowing for the deposition of the Chalky boulder clay, total (net) volume eroded from East Anglia over the past 450 ka is estimated at 629 km³. Of this total, 63% is attributed to normal erosion, 31% to the work of the Anglian ice advance, a mere 6% to marine erosion. Neglecting deposition of the Chalky till the work is more evenly divided; 56% of the total eroded is the work of ice, 40% the work of normal erosion, and just 3.9% the work of the sea. Considered separately, solution contributes only one fifth of the denudation achieved by normal erosion, 10% of gross, and 15.5% of net denudation.

Considering the till deposited by the Anglian advance, from a mean depth (Table 2) of 27.1 m (including the allowance for solutional lowering of the till surface), an average of 21.3 m has been removed (Table 3). Thus the mean depth of till remaining in areas C and D of Figure 3 is a little under 6 m; i.e. it appears that almost 80% of the Anglian till has been removed by the process of normal erosion over the last 440,000 years. However, depths close to the original still remain in many plateau areas, whilst elsewhere the erosion has cut into the underlying solid rocks, so the true figure must be about two-thirds lost. Though simple arithmetic might suggest that the rest could be lost in a further 120,000 years or so, the spatial irregularity of erosion, both past and present, implies that till fragments will survive on interfluves for at least another 440,000 years, provided only that a future ice sheet does not extend as far south as the Anglian limit.

Table 5: Overall summary and comparison of erosional work done

	Rate Bubnoff) (mm/10^3 yr)	Area affected (km^2)	Time of operation (kyr)	Total volume eroded (km^3)
Glaciation	3680	7,544	≈20	**555.3**
Deposition		13,368	≈20	**-304.1**
Glacial net	601	20,912	≈20	**251.2**
'Normal'	**48.4**	21,672	450	472
of which				
Solution	**~10**	21,672	450	97.5
Marine	3103	≈350	≈36	**~39.1**
Total	75	22,584	450	**762.3**
Allowance for solution of chalky till				
Glaciation	3680	7,544	≈20	**555.3**
Deposition		13,368	≈20	**-362.9**
Glacial net	460	20,912	≈20	**192.4**
Normal but excluding Fens				
'Normal'	**48.4**	18,248	450	397.4
New Total	61.8	22,584	450	**628.9**

Bold figures represent starting values from which other figures in the table are calculated.

Areas are from Figure 1.

Total area includes 1262 km^2 which is now sea, created by glacial erosion in The Wash and marine erosion along the cliffed coast of north-east Norfolk.

The last set of values allows for a) the allowance of 4.4 m for solution as an addition to measured thicknesses of Anglian till; b) the omission of the Fens from calculation of normal erosion given that it has primarily been an area of deposition since the Anglian glaciation.

COMPARISON WITH OTHER AREAS AND ERROR ESTIMATES

In one of the earlier papers already referred to (Clayton, 1997) a table of published rates for 'normal' erosion in the British Isles was included, reproduced here as Table 6. The limestone solution figures are included, partly because they are readily calculated from hardness values and because a large area of East Anglia is underlain by Chalk. In addition, the Carboniferous limestone is only a little way above the average resistance to erosion for all British rocks (Clayton & Shamoon, 1998b), so it may be regarded as representative.

It will be seen that the published values are consistent with the value of 48.4 B for normal erosion in East Anglia and with the rate for all processes of 61.8 B. Wider comparison is not perhaps appropriate here, but the interested reader is referred to the summary of worldwide rate of erosion values in Saunders & Young (1983).

Table 5 has figures in bold for the basic values from which others are calculated. If we take each in turn, we may estimate the possible errors which could affect them. For glacial erosion, the volume of 555.3 km^3 is probably correct within ±15%, as is the deposited till volume estimated at 304.1 km^3. The net work done by ice is the eroded volume less the till volume; the latter cannot exceed the former, and on the whole I believe the error may be limited to ±20%.

The normal erosion figure is derived from a mean depth of erosion for the area in Figure 4. In the original paper, errors were discussed, with the conclusion that the depth of erosion (and from that the rate) was unlikely to be overestimated, but that it could be underestimated by up to 15%. A further error could result if the area is not representative of the relief (and geology) of East Anglia. With the exception of the Fens (allowed for in Table 5) it seems very typical of the current and past relief and geology of the area considered here, and I see no reason to add to the error estimate of +15/-0% already given.

Solution is included within normal erosion, and derives from earlier work by Edwards (1973). The main error (other than securing a representative value for hardness) is how far current conditions may be held to represent the past. Edwards himself did his best to eliminate such errors, but even so it is probably right to allow for something of the order of ±20%.

Table 6: Published rates of denudation, British Isles

Area	**Reference**	**Rate (B)***
a) Reported rates of lowering for limestone solution, British Isles		
Southern Pennines	(Pitty, 1968)	75-83
Fergus Basin, Ireland	(Williams, 1963)	51
Mendip Hills	(Drew, 1974)	50-100
b) Reported rates of denudation by all processes, British Isles		
Bain valley, Lincs	(Straw, 1979)	92 ($)
Devon	(Walling & Webb, 1978)	9-125
River Ouse & Adur, Sussex	(Collins, 1981)	55-132
North Yorks Moors	(Arnett, 1979)	19-54
River Tyne, Northumberland	(Hall, 1967)	70-127

* The term Bubnoff, whilst not universally popular, has the advantage that it is numerically the same value as either the average rate of surface lowering expressed as $mm.10^{3} yr^{-1}$, as $m10^{6} yr^{-1}$ or the rate of removal of material expressed as $m^{3} yr^{-1} km^{-2}$.

$ Was erroneously reported by Saunders & Young (1983) as 124B. The value of 92B reflects Straw's assignment of the till to the Wolstonian; if it is of Anglian age, the rate would be 25B.

Marine erosion, although it appears in bold as a volume, is actually derived from the extrapolation over 6000 years of current rates of coastal retreat combined with the necessary adjustment of sea floor geometry to secure constant wave energy at the coast. The total is then estimated from three interglacials at 10 ka each plus the current interglacial estimated (in terms of the time since the end of the Flandrian transgression) at 6 ka. These time estimates are rather arbitrary and cannot be checked from current knowledge, especially of sea-level history within past interglacials. This suggests the marine erosion figure for the past 450 ka could be as much as +10/-40% in error. The

larger figure for overestimate allows for the fact that variations in sealevel will tend to reduce total erosion at the equivalent of the contemporary coastline. Fortunately this is the smallest contributor to total eroded volume, so the overall effect is not particularly serious.

Some small further errors may be involved in the estimation of rates from volumes, though this is greatly constrained by the fixed total area and the fixed timescale selected. Finally, in considering the precision of this work, every effort has been made to try and establish a 'most probable' figure for erosional wok done, and the broad conclusions and relative placing of the various styles of erosion is unlikely to be affected by adjustments which may be made to these values as further work on this topic is pursued.

CONCLUSION

In southernmost England, beyond the limit of the Anglian ice advance, denudation is necessarily dominated by normal erosion, with a small (<10%, even on relatively weak rocks) contribution from marine erosion. In northern England, Wales and Scotland, denudation over the past half million years is dominated by the work of ice (Clayton, 1996). East Anglia lies between, both geographically and in terms of the relative contributions of glaciation and normal denudation. Thus, while the overall rate of erosion is far lower than in the areas affected by multiple glaciation, normal erosion have been relatively unaffected by the work of the Anglian ice sheet, giving an overall rate of denudation here higher by that amount when compared with unglaciated Britain. The Anglian glaciation reduced the Fens to such a low level that normal erosion has been unable to achieve any measurable work since than (indeed the embayment has been filled by marine silts and peat growth). The till deposited around the glacially eroded basin has not only survived until today, but so long as no more ice sheets advance south of The Wash, it is likely to be still discoverable by future geologists when the glaciation is a million years in the past.

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[Manuscript received 10 June 1999; Revision accepted 21 February 2000]

CONTENTS

The Geological Society of Norfolk exists to promote the study and understanding of geology, particularly in East Anglia, and holds monthly meetings throughout the year.

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The illustration on the front cover is figure 6 from the article by Fish *et al*., in this issue of the Bulletin. It shows a possible explanation of 'erroneous' Chalk provenance data from tills at Weybourne.

ISSN0143-9286